HEAT BALANCE STUDIES AT THE CHILTON VALLEY, CASS IN THE NEW ZEALAND SOUTHERN ALPS

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ABSTRACT

This study shows that a knowledge of the values of the heat balance at the earth's surface gives an increased understanding of the character of the climate of an alpine valley.

Daily, monthly, and annual values of net radiation, soil heat flow, sensible heat flow and evaporative heat loss are measured, or estimated, for the period 15 August, 1969 to 14 August, 1970 at the Chilton Valley in the New Zealand Southern Alps (altitude 780 m.a.s.l.). In the estimation of these values use is made of a regression equation between net radiation and incoming shortwave radiation, non-weighing lysimeter measurements, and a modified form of the Penman formula for calculating daily values of evapotranspiration. Also reported are subsidiary studies of related phenomena such as the incident shortwave radiation on slopes, ice needle growth, the wetting and drying of soils, and advection.

The average values of the energy flows during the study year were:- net radiation, 136 ly day^{-1}, soil heat flow, 2 ly day^{-1}, sensible heat flow, -58 ly day^{-1}, and evaporative heat loss, -82ly day^{-1}. Spectral and harmonic analysis showed the daily values of the heat balance components to contain periodicities. The application of Lettau's theory of climatonomy illuminated other features of the climate. The daily energy flow values are shown to be interrelated, and to be associated with major synoptic weather types. In comparison with the heat balances of other mid-latitude

stations, near the west coasts of land masses, the heat balance of the Chilton Valley during the study year was noteworthy for (a) a positive mean monthly value of net radiation in winter, (b) a net mean monthly flow of sensible heat away from the surface in winter, and (c) the possibility of high monthly mean and daily values of the Bowen ratio, owing to soil moisture deficits in months of consistently high net radiation.

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LIST OF CONTENTS

LIST OF FIGURES

List of Figures (Contd.)

List of Figures (Contd.)

List of Figures (Contd.)

List of Figures (Contd.)

LIST OF PLATES

LIST OF TABLES

LIST OF SYMBOLS

Roman Capital Letters

A	Soil heat flow, constant in regression equation.
A_i	Amplitude of harmonic
A^o	Angström ratio
$A_{r,p}$	Areas of raingauge and lysimeter field tank orifices
AACC	Aperiodic advective contribution component
AET	Actual evapotranspiration
B	Inverse Bowen ratio, constant in regression equation
C	Heat capacity of soil
C.C.	Simple linear correlation coefficient
E	Evaporation of water
Ea	'Drying Power' term in Penman formulae
ET	Evapotranspiration
F	Forcing function in climatonomy
F'	Response function in climatonomy
$F_{1,2,3,}$	Soil heat flux plates
H	Kruskal-Wallis statistic
L	Latent heat of vapourisation of water
LE	Evaporative heat flow
LW	Longwave radiation
LW ↑	Longwave radiation from the surface
LW ↓	Longwave radiation towards the surface
M	Mean of time series values
M.C.C.	Multiple correlation coefficient
P	Sensible heat flow

Pe	Percolation
PET	Potential evapotranspiration
Q	Direct shortwave radiation
Ra	Precipitation
Rn	Net radiation
Ru	Runoff
S	Soil moisture storage
SW	Shortwave radiation
SW ↑	Shortwave radiation away from the surface
SW ↓	Shortwave radiation towards the surface
S.E.E.	Standard error of estimate in regression equations
T	Temperature, clear sky transmissivity
$V_{r,p}$	Volumes of water of precipitation and percolation
X	A variable

<u>Roman Lower Case Letters</u>

a	Exponent in wind power law, constant in regression equation
b	Constant in regression equation
c_p	Specific heat of air at constant pressure
dt/dz	Temperature (t) gradient with depth (height) (z)
e_a	Saturation vapour pressure at mean air temperature
e_d	Mean vapour pressure of the atmosphere
f (u)	Wind function in Penman formulae
i	An integer, number of harmonic
k	Coefficient of thermal conductivity, Von Karman's coefficient

n	Frequency of basic cycle
$n_i t$	Wave number
q	Diffuse shortwave radiation
u	Wind run, wind velocity
u_*	Friction velocity
w	Period
x_m	Volume fraction of mineral matter
x_o	Volume fraction of organic matter
x_w	Volume fraction of water
z	Height
z_o	Roughness length

<u>Greek Capital Letters</u>

β	Beta, partial impedance relating to LW↓
Γ	Gamma, partial impedance relating to LW↑
Δ	Delta, slope of the saturation vapour pressure curve at mean air temperature
ΔS	Change of soil moisture storage
Δ_i	Amplitude
Z_i	Zeta, total climatic impedance in climatonomy $(\Delta_i F / \Delta_i T_o)$
Σ	Sigma, sum
Φ	Phi, partial impedance relating to P
χ	Chi, partial impedance relating to LE
Ψ	Psi, partial impedance relating to A

<u>Greek Lower Case Letters</u>

α	Alpha, Albedo
b	Beta, phase constant relating to LW↓

γ	Gamma, phase constant relating to LW↑, psychrometric constant
δ_i	Delta, phase angle of forcing function
δ_i*	Phase angle of response function
ε	Epsilon, surface emissivity
ζ	Zeta, total phase angle in climatonomy (δ_i* - δ_i)
λ	Lambda, longwave exchange coefficient
μ	Mu, caloric admittance to the soil
π	Pi, a constant
ρ	Rho, density of air
σ	Sigma, Stéfan-Boltzmann constant, standard deviation
σ_p	Periodic dispersion in air temperature values
τ	Tau, period of basic cycle ($2\pi/n$)
φ	Phi, phase constant relating to P
χ	Chi, phase constant relating to LE
ψ	Psi, phase constant relating to A

CHAPTER ONE

INTRODUCTION

1.1 Heat Energy at the Surface of the Earth

Flows of heat energy into and out from the surface of the earth obey many laws of classical physics, the most important being the law of conservation of energy. It is this law that enables the heat balance or energy exchange at the earth's surface to be stated in one of its more simplified forms as

$$Rn - A - LE - P = 0. \qquad \text{—— } 1.1.1.$$

Here, Rn is the net radiant heat flow, A is the flow into the submedium and LE and P are the evaporative and sensible heat flows. This equation is valid both instantaneously and over any given time period. The above form omits some terms which are usually small in magnitude.

In physical climatology a great deal of attention has been given to the study of energy exchanges. An important reason for this is that the exchanges directly influence many other aspects of the atmosphere and submedium. In other fields, such as ecology, glaciology and hydrology, heat balance studies are made for the same reason.

Energy exchange studies may use any time or spatial scale. For example, Budyko (1958) has constructed world maps of annual values of the heat balance components, whereas Frankenberger (1962) has reported daily values of the components for one site in Germany. Heat balance studies also vary in the degree of direct measurement involved. Swinbank and Dyer (1968) have directly measured all the components. On the other hand, climatonomy theory (Lettau, 1968) enables heat balance component values to be calculated with the direct measurement of only downward shortwave radiation, providing certain background information is available.

Most of the major types of climate of the world have been examined to some extent with the use of the heat balance approach. However, the more severe types of climate have been given less attention. The climates of highland and mountain areas are a case in point. In mountain climates it is common for energy exchanges to be measured only for short periods, as, for example, in the work of Terjung et.al., (1969). This is mainly due to the practical difficulties of maintaining instrumentation for extended lengths of time. Studies where daily heat balance component values are obtained for a period of a year, such as those of Wendler (1970), are rare. The present study represents a contribution to the knowledge of the heat balance climatology of mountain areas in general. In particular, the heat balance approach is used to give insight into the climate of a New Zealand alpine valley.

1.2 Aims and Structure

The environment chosen for study is that of the Chilton Valley, latitude 43° 02' S longitude 171° 46' E, located near Cass in the Southern Alps of New Zealand. The principal aim of this study is to demonstrate that a knowledge of climatic energy exchanges gives an insight into the character of the climate of this environment. In order to achieve this aim three procedures are used. These are:-

(1) The measurement or calculation of daily and annual values of the heat balance.

(2) The analysis of the heat balance values.

(3) The completion of selected subsidiary studies that are necessary for the heat balance calculations, and/or show the relation of energy exchanges to other features of the climate of the location.

The integration of these procedures in the structure of the study is diagrammatically shown in Figure 1.1. Background information of three kinds is necessary. Firstly, relevant, non-climatic features of the physical geography of the valley are described. Then a standard climatic description is presented. This may be used, in part, as a basis from which the increased understanding, attendant upon the use of the heat balance approach, may be measured. Finally, details of the instrumentation, periods of data collection, and accuracy of the heat balance estimates are given. This background information forms the rest of this chapter.

Chapters 2 to 5 concentrate on the measurement or calculation of the four main heat balance components, Rn, A, LE and P, and are in accord with the first procedure listed above. Although the magnitudes of the energy flows are central to the whole study, these chapters also include reports of the subsidiary studies mentioned in the third procedure. In Chapter 2, for example, the calculation of Rn values cannot be made without information on other component parts of the

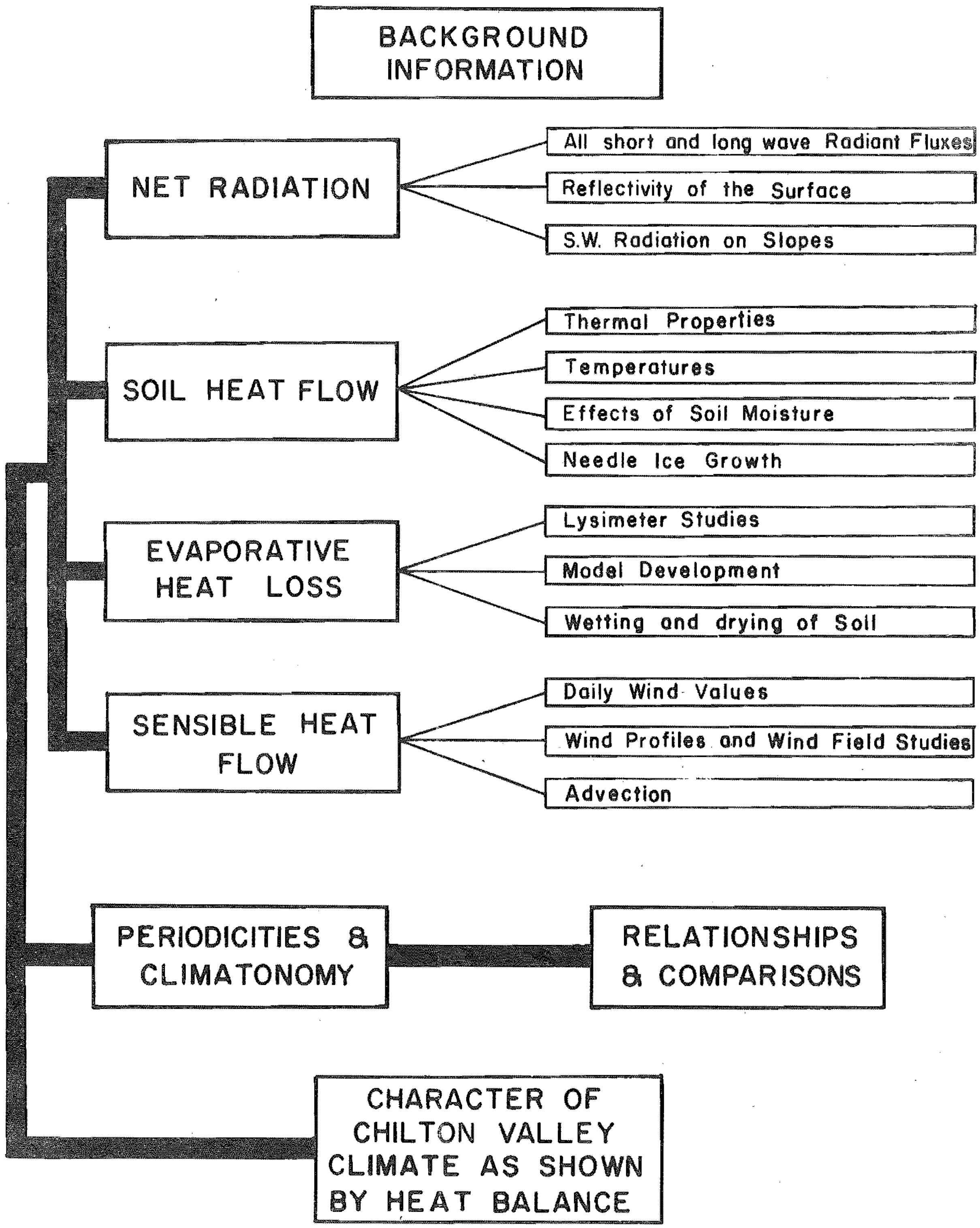


FIGURE 1.1: The structure of the present study

radiation balance. The Rn values themselves, are then supplemented by an estimation of SW↓ at other locations within the valley. Similarly, subsidiary studies are reported in Chapters 3 and 5 where, respectively, values of A and P are germane to the study as a whole. In Chapter 4, the calculation of LE necessitates a full discussion of the method of computation of daily values of E. It should be noted that the subjects of the subsidiary studies are not related exclusively to the energy flow indicated in Figure 1.1, but in many cases are also related among themselves.

Chapters 6 and 7 are concerned with the analysis of the daily and annual heat balance values, which is the second procedure noted above. In Chapter 6, the daily heat balance values, when analysed as a time series, are shown to contain periodicities. The presence of these periodicities is emphasised by the application of climatonomic theory. This application, in turn, draws attention to other features of the climate of the valley. In Chapter 7, the interrelationships of the daily energy flows, and the effect of the synoptic weather on their values are examined. Also, a comparison with heat balance estimates for other parts of the world is made.

In Chapter 8 the main observations and conclusions from earlier chapters, that are pertinent to the principal aim of the study, are reviewed within the context of this aim.

Finally, comments on the extension of the results from the study and suggestions for further research are given.

Overall, (Figure 1.1), the study proceeds through the derivation of the heat balance component values (left hand side of the diagram), with relevant subsidiary studies at appropriate points (right hand side of the diagram). Then, an analysis of the energy flows is made (lower part of the diagram). Finally, (base of the diagram) a synthesis of the major findings is given to demonstrate the value of the heat balance approach as an aid in understanding the climatic characteristics of the valley.

1.3 The Chilton Valley

The Chilton Valley (Plate 1) is a small valley on the flanks of Sugarloaf Hill in the Cass Basin (Figure 1.2). The basin is part of the Waimakariri catchment which Hayward (1967) describes as a high altitude, mountain catchment with predominantly steep, unstable slopes, and shallow, infertile and erodible soils. The vegetation has been drastically modified in pre-European and European times, and consequently, half the catchment is in a severely eroded condition. The mean altitude of the catchment is approximately 1065 m. The catchment as a whole is, at present, in a delicate state of equilibrium. Whereas careful land management practices may help to decelerate erosion, any cataclysmic event such as an earthquake, fire or storm, might well set off a new cycle of erosion. As little can be done about such events, concentration on the 'normal' physical and biological processes within the catchment becomes important. With greater knowledge of these processes, and the application of this knowledge through land management practices, the catchment might be made progressively more stable.

The Chilton Valley itself is typical of what Hayward (1967) calls the central montane (305-925 m) part of the

PLATE 1: The Chilton Valley, Cass

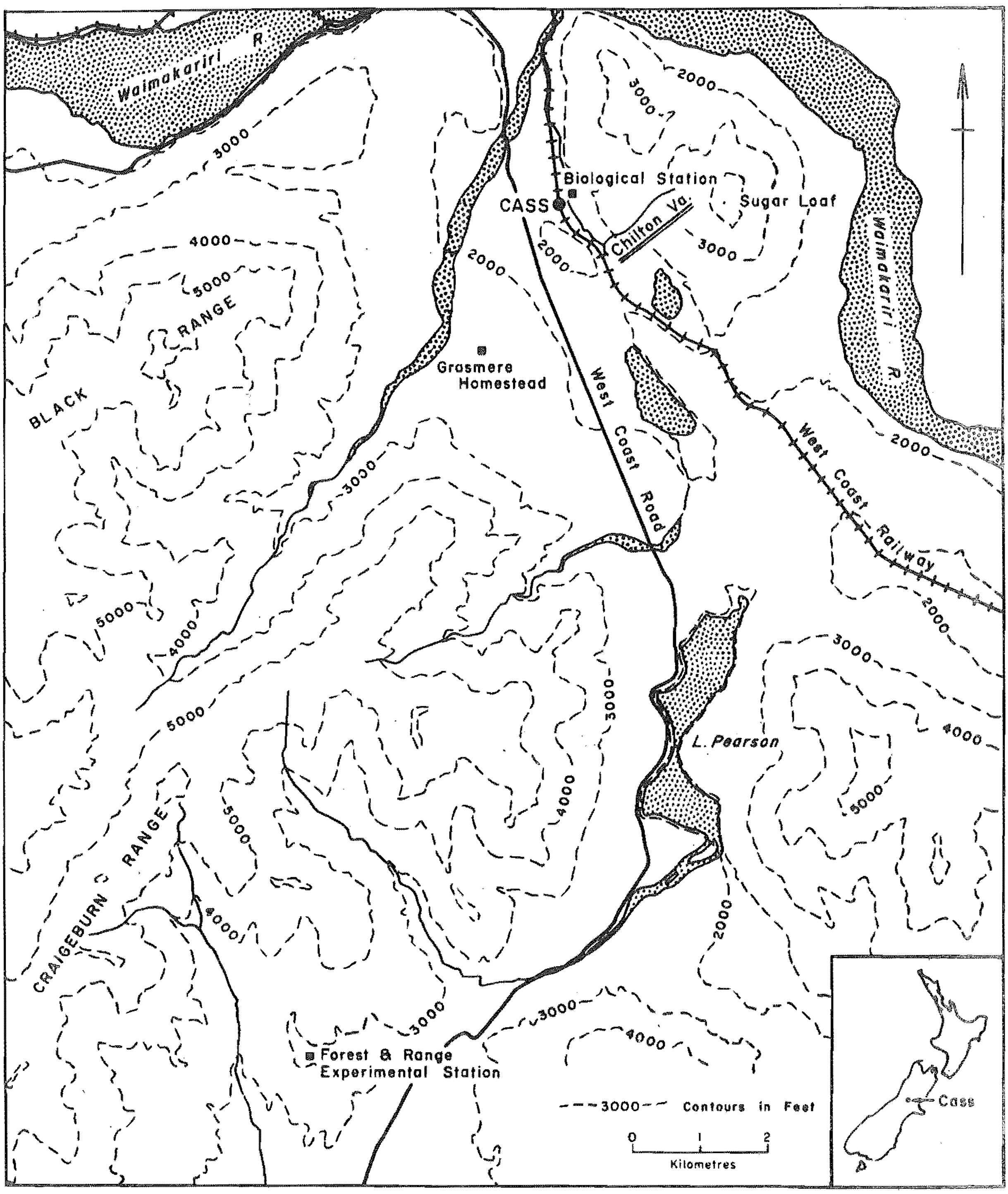


FIGURE 1.2: The location of the Chilton Valley

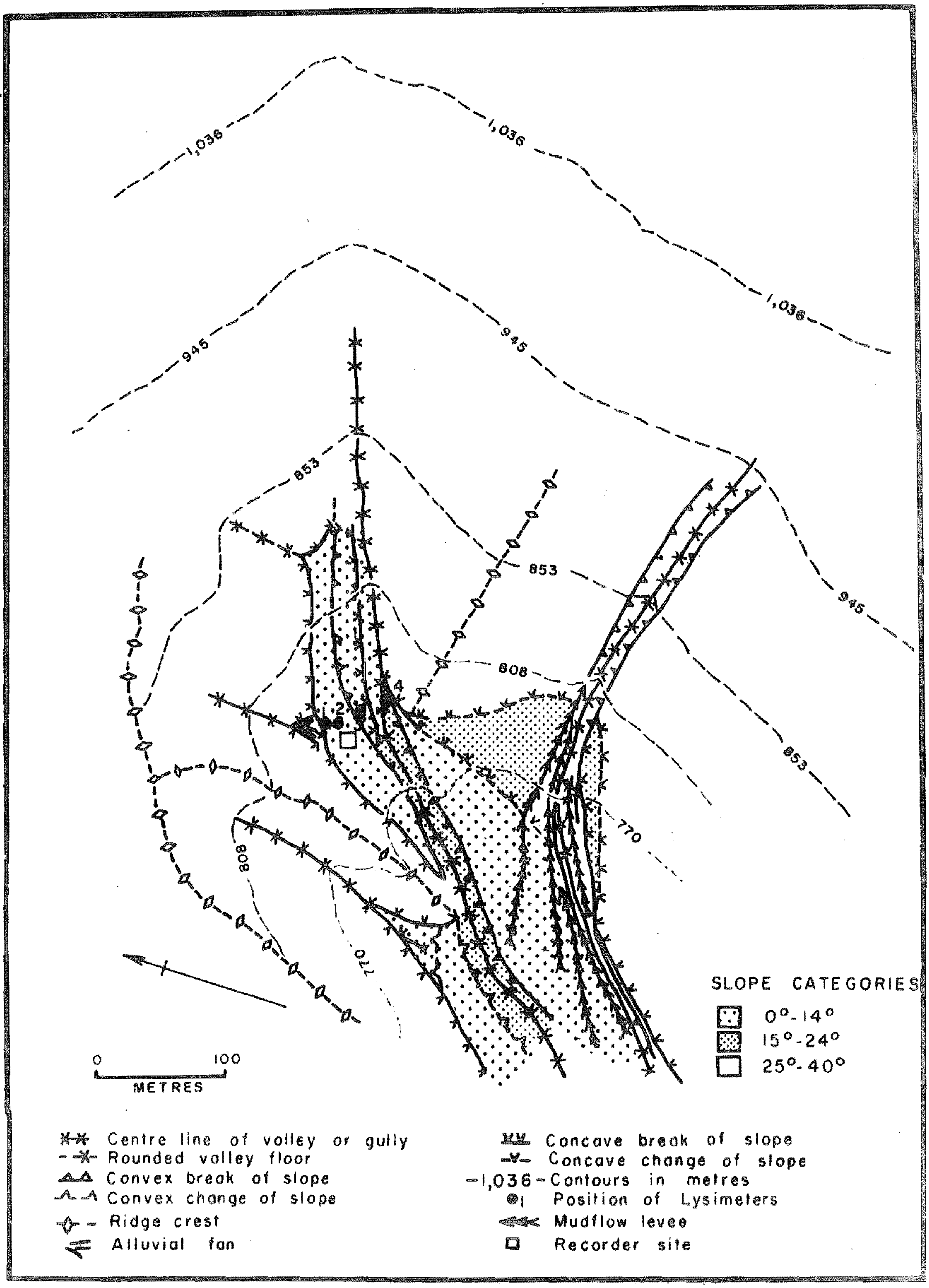


FIGURE 1.3: The topography of the Chilton Valley

Waimakariri catchment. The valley is aligned northeast-southwest, and rises from about 700 m at its mouth, the top of an alluvial fan, to 1360 m at Sugarloaf summit. The main recording site is at 780 m. The valley has a large range of slopes and aspects (Figure 1.3). Most slopes are mantled by periglacial screes of unknown depths.

The soils of the valley are upland and high country earths of the Kaikoura Series. The modal profile of this greywacke-derived soil consists of 13 cm of very dark, grey-brown, crumb, very friable sand loam. This overlies 30 cm pale, yellow-brown, gritty, friable, silt loam on pale, yellow-brown, very stony, loose, sandy loam (Hayward, 1967). Further details have been given by McDonald (1961).

Variations in vegetation (Plates 1-8) are influenced by aspect. High on the north-west facing slopes there are extensive bare areas, while lower down, there is a large area of manuka (*Leptospermum scoparium*) with a few scattered beech trees (*Nothofagus solandri var. cliffordioides*) and smaller scree patches. The valley floor is mainly covered with hard and silver tussock (*Festuca novae-zelandiae* and *Poa caespitosa*) together with relatively low scrub of cassinia, matagouri, snow totara, and hebe (*Cassinia fulvida*, *Discaria toumatou*, *Podocarpus nivalis*, and *Hebe buxifolia*). On the south and south-east facing slopes, the tussocks are accompanied by mountain daisies (*Celmisia spectabilis*) and inaka (*Dracophyllum uniflorum*). The vegetation cover at the main recording site is shown in Figure 1.4. Further details on the vegetation and other physical features of the valley have been given by Soons and Rayner (1968).

1.4 The Climate of the Chilton Valley

Standard descriptions of New Zealand High Country climates have been made for several locations (Morris, 1965, Coulter, 1967, 1969, Cunningham and Gannaway, 1969). A detailed account of the climate of the Chilton Valley in terms of standard climatic parameters will furnish information comparable with these earlier descriptions. However, the main aim of this section is to provide a starting point in climatic description. It is then proposed to show, in the remainder of the study, how a knowledge of the energy flows increases our understanding of the climate of this location.

The synoptic scale weather of the South Island, that forms the context of the climate of the Chilton Valley, is dominated by an almost regular procession of eastward moving anticyclones. The time between successive high pressure systems is about a week (de Lisle, 1969). Between each pair of anticyclones there is a trough which often contains a cold front. Fuller details have been given by many authors e.g. Maunder (1970). The South Island mountain chain influences this large scale pattern by sometimes retarding the passage of a front, and often creating a low pressure area to the leeward, over the Canterbury Plains (Sevelle, 1969). Föhn wind effects are also a frequent occurrence (Lamb, 1970).

The topoclimate of the Chilton Valley, while dominated by the passage of the major synoptic systems, has many of the characteristics of mountain climates, as described by Coulter (1967). Although the valley tends to be sheltered from the highest winds, the effects of orographic rain from the west, relatively low temperatures, with high diurnal extremes, and temperature inversions are common. Specific details of the local climate will now be presented. This account is based on climatic data collected at the site from 1964 up to August 1970.

Records of air temperature at 3 m are presented with data from the University of Canterbury Biological Field Station (C.J. Burrows, pers.comm.), and the Craigieburn Experimental Station of the New Zealand Forest Service (Morris,1965) (Table 1.1). Temperatures in these stations are taken in Stevenson screens that are nearer the surface than 3 m. Only the Craigieburn records are compiled from data that are complete for each month. Despite these inconsistencies, and the three sets of data being from different years, the comparison shows a general similarity both in annual means and ranges. As expected, all the annual means are lower than the 10.7^{o}C value at Christchurch Airport for 1964-66 (N.Z.M.S. 1966-68). A decrease of annual mean temperature with height may be seen between the three stations. In addition, and probably due to cold air drainage effects, mean temperatures in July and August are higher at the Chilton Valley site than at the Biology Station. However, confirmation of this situation must await a more specific study of these effects. Average monthly maximum and minimum temperatures in the Chilton Valley (Table 1.2) indicate the relatively high temperatures reached in summer, and the occurrence of frosts in winter. Absolute maximum and minimum air temperatures that have been recorded are 40^{o}C and -8.5^{o}C respectively.

Field work in January 1967, showed that the valley is above the effect of extreme inversions. Comparison over

TABLE 1.1

COMPARISON OF MEAN AIR TEMPERATURES IN ^{o}C
IN THE CASS DISTRICT

Station	Chilton Valley	Botany Hut	Craigieburn (Forest)
Altitude m	780	566	808
Approx. distance from Chilton Valley. km	-	1	13
Years of Record	1964-68	1961-1964	1961-63
January	14.3	14.6	14.5
February	16.0	15.7	14.5
March	12.5	13.2	11.2
April	10.5	8.8	8.8
May	5.5	5.8	5.6
June	2.5	2.7	1.7
July	1.7	1.6	1.1
August	4.0	3.7	2.2
September	6.5	5.7	5.0
October	8.0	9.8	9.4
November	8.2	11.5	9.4
December	13.5	15.5	13.9
Year	8.6	9.0	8.3

TABLE 1.2

MAXIMUM AND MINIMUM AIR TEMPERATURES IN °C
FOR THE CHILTON VALLEY 1964-68

Month	Maximum	Minimum
January	18.5	8.8
February	20.2	10.5
March	19.5	10.2
April	14.5	5.5
May	9.0	1.7
June	6.5	-1.0
July	5.7	-2.7
August	8.0	0.5
September	10.2	2.2
October	13.5	4.4
November	14.0	4.8
December	18.0	9.5

the altitude difference (214 m) between the recording site and the Biology Station indicated that, in this month, inversions in the main basin occurred 40% of the time, with the greatest temperature difference being 6.1°C. In winter, however, inversions of 10°C temperature difference are frequently recorded. Within the valley itself, records between the recording site and the top hill thermistor (888 m) showed inversions to occur 31% of the time. Furthermore, air temperatures between 12 m, at the top of the mast, and screen height, showed local inversions to occur on 57% of the days of the study period defined in the following section. In this period valley inversions, between the hill thermistor and the recording site, were most frequent in November, May and June but occurred all the year round (Table 5.7).

Precipitation records have been kept at the Biological Station since 1918. Up to and including 1965 the average annual rainfall was 131 cm (51.2 ins.). The variability, as defined by Seelye (1940), of 14.91% is relatively high, and emphasises the fact that the location is in an area of steep annual rainfall gradient, between high values in the west and low values in the east. There is only a small variation within the year, ranging from 8.9 cm in February to 13.0 cm in October, with spring showing a tendency to be the wettest season, and late summer the driest. There is a close relationship between the Biology Station rainfall and the Chilton Valley records (Greenland and Owens, 1967). These authors also showed that wet and dry periods tend to occur in groups of two days or longer, with fewer than one third of the wet and dry periods occurring as single days. This persistence, largely due to synoptic conditions, is important especially in the process of wetting soils, and also, to a certain extent, in their drying. However, the wet periods do not have high rainfalls, half of them being less than 1.5 cm in total. Rainfall intensities are also low. Maximum 24 hour intensities for two and 20 year return periods, estimated from the work of Robertson (1963), are 8.1 and 16.5 cm respectively. Little is known about precipitation in forms other than rain. Soons and Rayner (1968) report that in the Cass Basin, snow may lie at low levels for a day or two on a few occasions in most winters, but it is unusual for it to persist for any length of time.

Wind velocities, from approximately two years' data at the Biological Station (C.J. Burrows, pers.comm.), indicate an average wind speed of 4.9 $m.sec^{-1}$., when corrected for instrumental error (see section 5.3). The highest and lowest average wind velocities were recorded in spring and winter respectively. Owing to greater shelter, the average velocities in the Chilton Valley, not measured until the study year, would be expected to be lower than the above figure. Wind directions at the Biological Station show that winds from a north-westerly quarter are the most frequent, occurring 50% of the time. Percentage frequencies for other directions are:- south east to north east 16%, south to south west 13%, calm 20%. In the valley, owing to its topography, north west winds are transformed to north east winds, and winds with a southerly component are transformed to south easterly winds.

No humidity data are recorded at either the Chilton

Valley or the Biological Station. The average relative humidity at Craigieburn (1964-66) was 74% (N.Z.M.S. 1966-70), and at this site values of 20-30% are often recorded during north-westerly air flow (Hayward, 1967). Owing to the relative proximity of the Craigieburn site, similar humidity conditions might be expected at the Chilton Valley.

A seasonal analysis of incoming shortwave radiation, from December 1964 to December 1967 (Table 1.3) shows 1965/66 to have the highest radiation of the three years. Autumn has the highest percentage of possible radiation in all three years. Monthly values (Table 1.4) indicate this feature in more detail, with May, in particular, receiving over 80% of the radiation possible at the site under clear sky conditions. In contrast, spring, and November in particular, receives the lowest amounts of radiation. If it is assumed that there is an association between cloud cover and precipitation, then there is general agreement with the annual rainfall distribution. This would indicate that the relatively short data period may be representative of 'normal' conditions.

Except for the inclusion of shortwave radiation information, the above climatic description is similar, in many ways, to other reports of New Zealand High Country climates. With the aid of the heat balance approach it is the purpose of the present work to add insight into the character of the climate described above.

1.5 Instrumentation, Data Collection and Accuracy

The environment of the Chilton Valley is relatively severe. In order to measure or calculate the heat balance on a daily

TABLE 1.3

SEASONAL TOTALS OF RADIATION IN KILOLANGLEYS

(Figures in Brackets Indicate the Percentage of The Total Possible Radiation)

Season	Max.poss.	1964/5	1965/6	1966/7	3 Year Average
Summer Dec 1 - Feb 28	60.62	42.32 (69.9)	43.40 (71.7)	40.50 (66.9)	42.27 (70.0)
Autumn Mar 1 - May 31	24.97	18.75 (75.2)	19.69 (78.9)	18.22 (72.8)	18.89 (75.8)
Winter June 1 - Aug 31	13.30	8.84 (66.5)	9.59 (72.0)	8.22 (61.0)	8.88 (65.1)
Spring Sept 1 - Nov 30	49.12	32.00 (65.2)	31.22 (63.7)	29.02 (59.1)	30.75 (62.8)
Annual Total	148.02	101.91 (69.0)	103.89 (70.2)	95.95 (64.9)	100.45 (68.0)

TABLE 1.4

MONTHLY MEAN TOTAL RADIATION IN LANGLEYS PER DAY AT THE CHILTON VALLEY SITE

Month	Max. poss. at recorder Site	1965	1966	1967	3 Year Average	% of Max. poss.
January	696	446	485	402	444	63.8
February	586	439	378	500	439	74.9
March	424	316	352	286	318	75.0
April	252	188	183	193	188	74.5
May	135	107	105	115	109	80.8
June	74	55	44	55	51	68.9
July	115	83	87	78	83	72.1
August	251	150	180	134	155	61.7
September	429	275	266	248	263	61.2
October	558	348	364	364	358	64.2
November	684	433	399	343	392	57.2
December	730	574	453	497	508	69.4

basis at this location, a combination of two kinds of instrumentation was used. Firstly, durable sensing and automatic recording systems were employed. Secondly, data, thus derived, were supplemented by means of instruments used on a shorter term or non automatically recording basis. The locations of the various instruments used at the main recording site are shown in Figure 1.4 and Plate 2.

Shortwave direct and diffuse solar radiation was measured using an Epply 16 junction pyranometer. Net radiation was measured using a Funk C.S.I.R.O. net radiometer. The latter was not used continuously owing to its susceptibility to damage by birds and opossums. Heat flux in the soil was given by three C.S.I.R.O. soil heat flux plates placed at 1 cm depth in the soil. Temperatures were measured by stable thermistors produced by the Yellow Springs Instrument Co. Of the thermistors used in this study, one was placed in a standard Stevenson screen, and four were in the soil at depths of 0.5, 2.0, 15.0 and 30.0 cm, in a profile about 70 cm from one of the heat flux plates. A further two thermistors were in radiation shields on a mast at heights of 3 and 12 m, while another was placed in a radiation shield 2 m above the ground, at an altitude of 888 m, 108 m above the main recording site. Signals from all of these transducers were recorded on a Honeywell-Brown 24 pt 1mV 2½ sec. full scan, potentiometric recorder. This was connected to a timing mechanism which allowed recordings to be taken at a normal frequency of once every 7½ minutes or else continuously, if required. The recording system incorporated an auxiliary battery driven power unit, in an attempt to counter frequent failures of

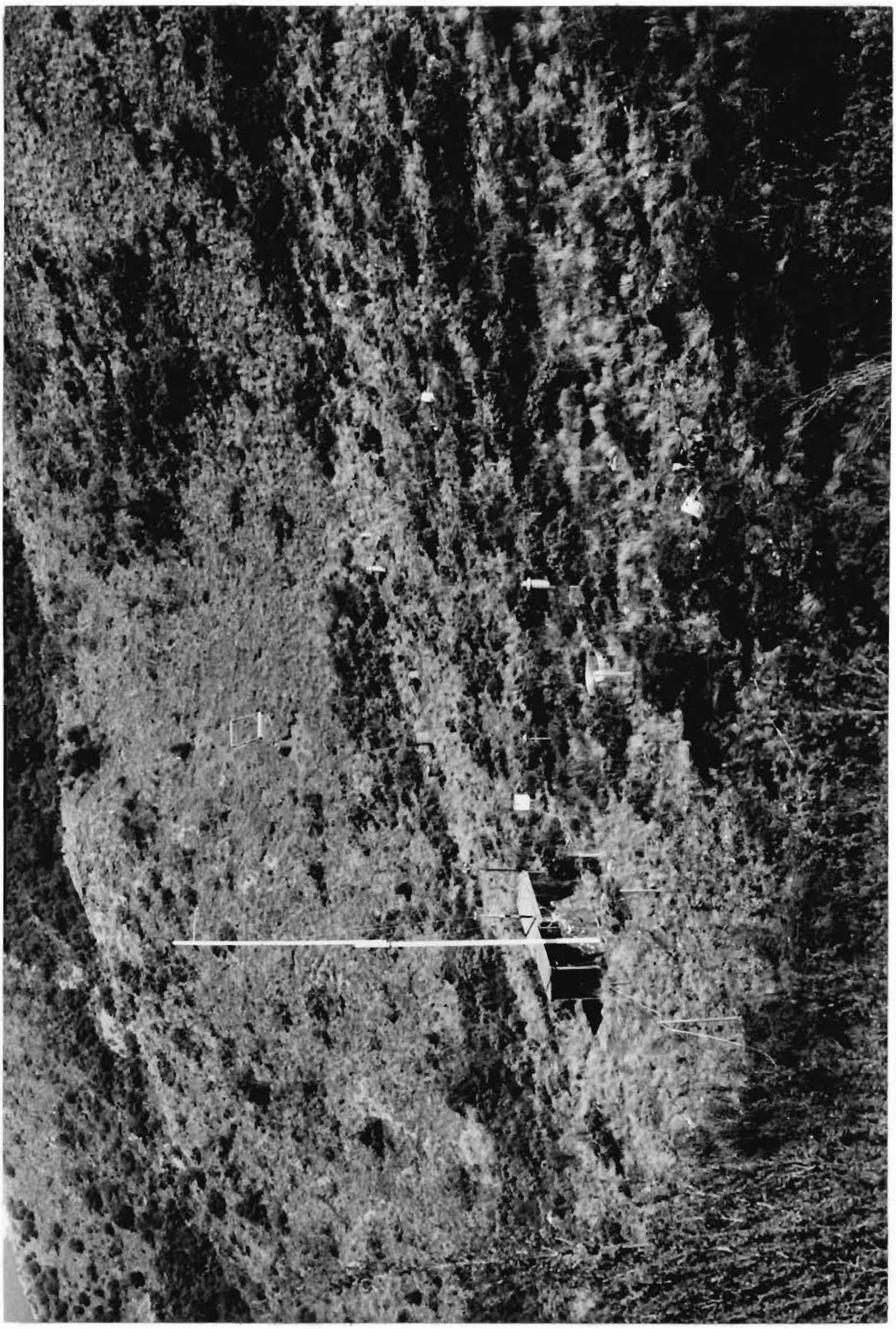


PLATE 2: The main recording site in the Chilton Valley

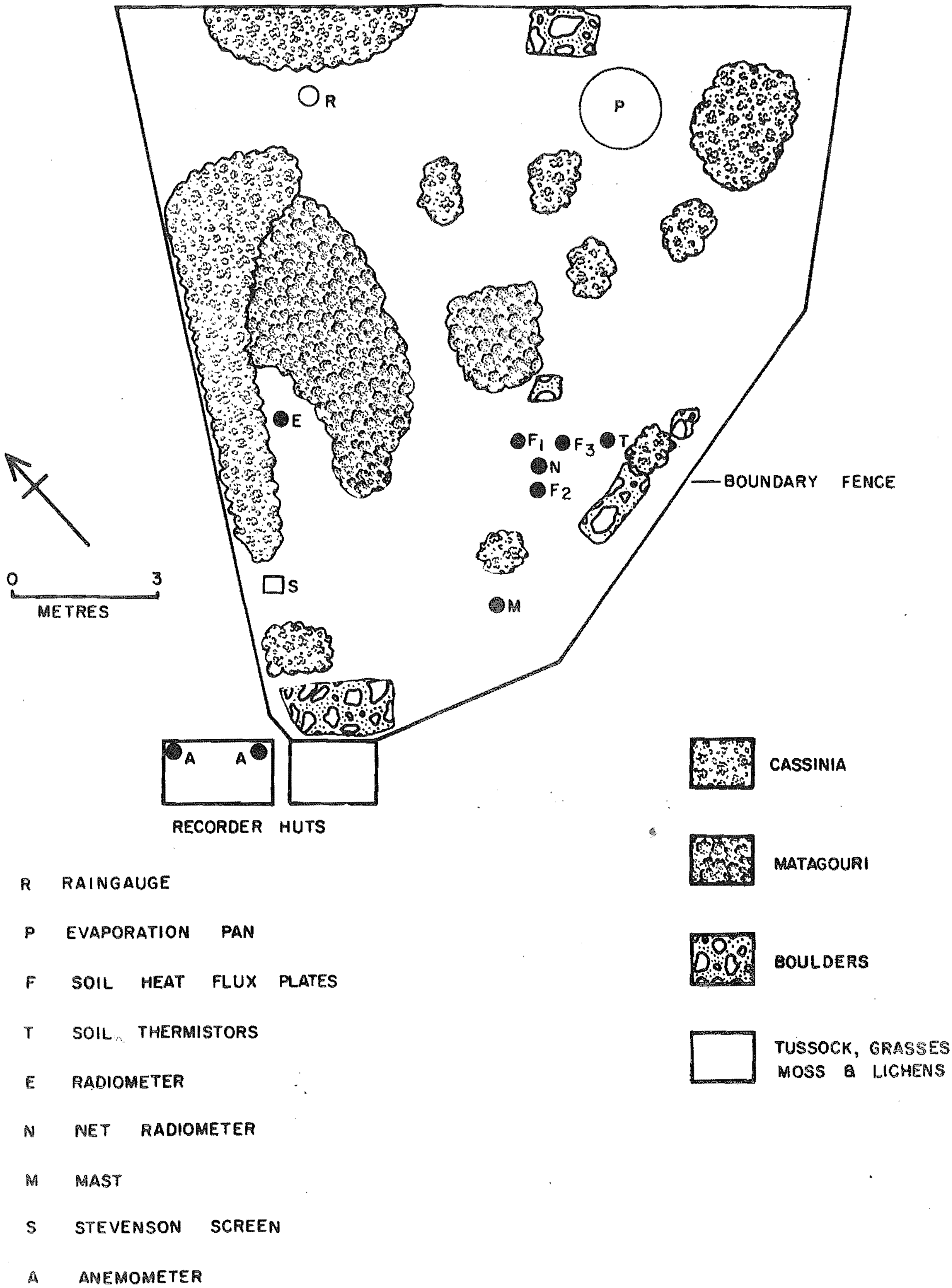


FIGURE 1.4: Vegetation and instrumentation at the main recording site

the power supply. Loss of records could be due to the failure of any one part of the whole system. Since, up until the year used in this study, visits to the valley were made only at monthly intervals, a long term continuous record of all variables was sometimes interrupted. Within the study year continuity was achieved with the exceptions stated below.

Besides this main recording system, wind was measured with a Lambrecht continuous velocity and direction wind recorder. The sensing head was located at 3 m above the surface. This record was interrupted when the velocity pen ink supply was exhausted. Short term wind velocity records were also taken with a Casella counting cup anemometer at the same height. Three OTA Keiki Seisakusho hand anemometers were also used.

Precipitation was monitored by a Lambrecht continuous rain gauge. Gypsum blocks and gravimetric methods were used to determine soil moisture. Actual evapotranspiration was measured directly by four non-weighing lysimeters placed across the valley near the main site (Plate 3). The lysimeters are described in section 4.2. In addition, a runoff plot 290 cms by 152 cm was converted, so that it gave continuous measurements of rainfall and percolation when in operation.

During intensive study periods, humidity was measured by an Assmann aspirated psychrometer, evaporation by a Ministry of Works evaporation pan, based on the design of the raised U.S. class A pan, and albedo was recorded using a C.S.I.R.O. albedometer.

In this study the main body of data is for the year

15 August 1969 to 14 August 1970. This period is referred to in the rest of this work as the study period. Data external to this period are used in some specified cases. During the study period, every effort was made to ensure a complete record, but even so there were breaks in the recording system, and times when the wind recorder lacked ink. These times are listed in Appendix A. The regression equations, between Chilton Valley and Craigieburn data, used to obtain values for days with missing data in the former location are also given in the same appendix.

An attempt has been made to estimate the likely error of the values of the heat balance components derived in this study. Budyko (1958) has suggested three ways of evaluating these; namely, comparisons of the results of various independent methods for calculating the components of the heat balance, comparison of calculations of the heat balance components to direct measurements taken with special instruments, and thirdly, by calculation of the error of all the components by closing the heat balance equation. Difficulties arise in the present study as the third method cannot be used at all, as P is taken as a residual, and the other two methods can be only partially used. Furthermore, errors of different kinds, instrumental, sampling and analytical, are present in some of the derived values.

At several points in the succeeding report errors in parameters, other than the main heat balance components, are discussed as they arise. In order to obtain estimates of error for the main heat balance components, two simplifying assumptions have been made. Firstly, it is assumed that errors can be computed with respect to the daily mean value of the component averaged over a year. One of the important effects of this assumption is the suppression of the fact that, in some cases, winter data are less accurate than summer data. As an example, the standard error of estimate of Rn, of 32 by day^{-1}, represents a far smaller relative error for the normally high summer values than that for the usually low winter values. The second assumption is that all of the errors dealt with are random in nature. This cannot be absolutely verified for all of the errors involved, but it is necessary where the errors have to be combined using standard techniques (e.g. Topping, 1955). Certain further assumptions are also made in the computations and are listed in Appendix B.

Estimates of percentage errors in the values of the heat balance components (Table 1.5) have been made using the above assumptions and in the manner shown in Appendix B. The error in Rn for annual and monthly values may be rather low owing to the assumption made concerning the standard deviation of climatological measurements, but in general, monthly and annual values compare well with those computed for calculations made by Budyko (1958). Estimates of errors for daily values are frequently omitted in the literature. The estimates shown in Table 1.5 indicate much higher errors than in careful work performed in some easily accessible experimental sites e.g. Frankenberger (1962). However, even in such conditions estimated errors of daily values can be high, as in the case of Högström's (1968) estimate of evaporation for which he quotes 10-15%. The effect of obtaining P by the residual

TABLE 1.5

ESTIMATED PERCENTAGE ERRORS IN VALUES OF THE HEAT BALANCE COMPONENTS

Time Period	Rn	A	LE	P
Year	1.3	*	10.0	10.1
Month	4.4	*	5.0	6.9
Day	23.8	5.0-10.0	5.8	29.7

* Error unable to be estimated. See Appendix B

method can be seen to lead to its value having the greatest error of all the components over each time period.

Other sources of heat flux which would be included in a complete energy balance for the earth's surface are ignored as their size is usually well below the accuracy level of the present measurements. These include:-

(1) Heat involved in the melting or freezing of snow, ice, or soil moisture. Some of the effects of this are reported in sections 3.5 and 3.6, but an estimate of this heat cannot be made on a daily basis for the study period due to lack of data. Owing to the low frequency of snow cover, and the small depth to which the soil may be frozen, neglect of this term is believed to constitute a small error compared with the errors in the main heat balance terms in the winter season.

(2) Heat used or generated by photosynthesis or respiration of plants. This is also relatively small. Photosynthesis on the average represents about 1% of SW↓ (Sellers, 1965) and in the present and similar locations here the vegetation growth rates are small (Mark, 1965; Terjung et.al., 1969) the use may be less.

(3) Vertical transport of heat by raindrops. This is also ignored although it is shown in section 3.5 that this can have a notable effect in winter.

(4) Heat from the frictional dissipation of the energy of the wind. This is estimated to lie between 1 and 10 ly day^{-1}. (Sellers, 1965)

(5) Flux of heat from the earth's interior. This has been found to average only 0.1 ly day^{-1} (Lee and MacDonald, 1963).

(6) Heat of condensation of water vapour. This is small when the error of evaporative heat loss measurement is considered.

1.6 Summary

In this chapter the principal aim of the work has been introduced and the main procedures used in the study have been outlined. Background information on the physical features and the climate of the Chilton Valley has been presented. The instrumentation and likely errors in the main heat balance components have been discussed. Attention is now turned to the measurement or calculation of the energy exchanges in this mountain environment.

CHAPTER TWO

RADIATIVE HEAT FLOW

2.1 Introduction

The ultimate energy source for virtually all climatological processes is the sun. However, direct solar radiation is but one of several radiant energy flows that arrive at the surface of the earth. The combination of all of the radiant flows arriving and departing from the surface is called the radiation balance or net radiation. The latter is defined by the equation

$$Rn = SW\downarrow (1-\alpha) + LW\downarrow - LW\uparrow \qquad \text{\_\_\_\_ } 2.1.1.$$

where $SW\downarrow$ is the incoming shortwave radiation and is composed of direct, Q, and diffuse, q, radiation. α is the albedo or reflectivity of the surface and $LW\downarrow$ and $LW\uparrow$ are the longwave radiation flows from atmospheric constituents and from the earth's surface, respectively. Net radiation is one of the energy balance components with which this study is concerned. In many senses it is the most important of the four principal components. For example, although there are exceptions, it generally plays the role of an energy source while the other components are energy sinks. The salient characteristics of Rn may be examined by describing the radiation balance itself and by investigating as many as possible of the terms of equation 2.1.1.

In this chapter, therefore, SW↓ is discussed and the results of albedo measurements are reported. It is then possible to obtain the net shortwave radiation. Rn itself, is calculated from SW↓. Since Rn is fundamental to the whole study, the relationship between it and SW↓ is examined in some detail. With data on Rn it is then possible to compute the longwave radiation balance. Consideration is given to the interaction of all of the radiation balance components. Finally, attention is turned to the amounts of SW↓ received on other slopes in the valley, as this is a logical progression from measuring the energy input at one point, and is also important in helping to explain some of the other physical phenomena occurring in the valley.

2.2 Incoming Shortwave Radiation

All radiation values refer to New Zealand Standard Time unless otherwise stated. Values of SW↓ for the study period (Tables 2.1 and 2.2) may be compared with the values for the years 1965 to 1967 (Tables 1.3 and 1.4). However, before discussing the values shown in these tables, it is necessary to explain how they may be adjusted in an attempt to remove the influence of sky line obstructions in the valley.

The recorder site in the Chilton Valley is surrounded by high ground giving rise to considerable shading, (Fig. 2.1). In order to compare the shortwave radiation recorded in the valley with that received in other locations, an effort was made to adjust the records so that they would represent those of a site, in the same position, with no sky line obstruction. This can be done for sunshine records using the method of

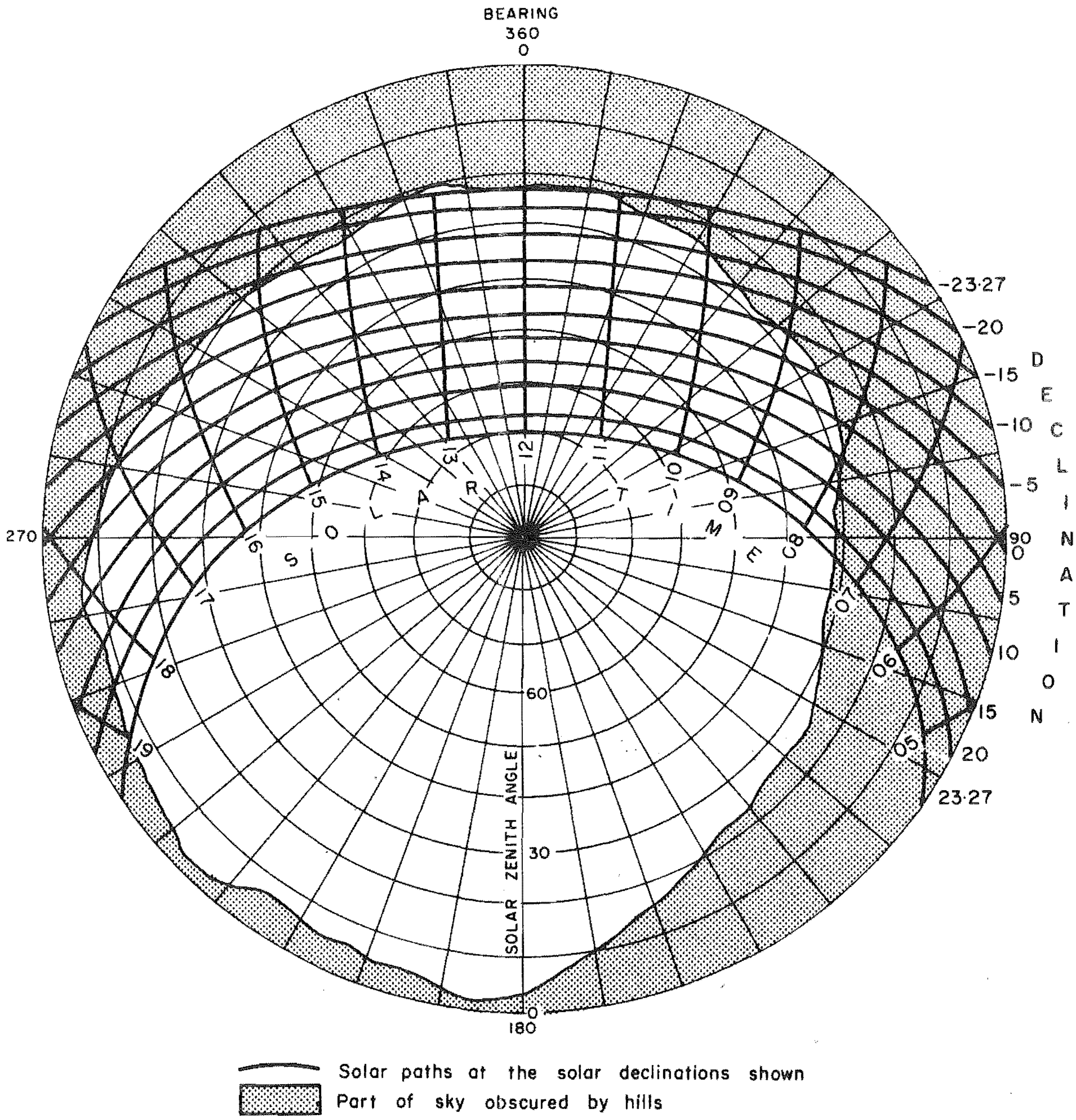


FIGURE 2.1: Solar path and horizon diagram for the recorder site

Furmage (1970), but for radiation values the work of Garnier and Ohmura (1968) can be applied. The latter have developed a method of computing the direct shortwave solar radiation on slopes which involves an integration, throughout the daylight period, of incoming direct shortwave radiation. A computer program giving hourly totals using their method has been written by Fuggle (1970). This was used to give hourly totals for the Chilton Valley site for eleven values of solar declination throughout the year using a slope value of 0.0 degrees and an atmospheric transmissivity of 0.75. Computations were made in local apparent time as is used in Fig. 2.1. Times of radiation loss due to sky line obstruction (Table 2.3) were taken from Fig. 2.1 and the actual direct radiation lost during these times was found from the computed hourly radiation totals. The figures for percentage lost were then used as a basis to adjust the monthly totals (Table 2.4). The method used does not allow for any change in diffuse radiation but, as is shown later in this section, this represents a fairly constant proportion of direct radiation at all times except winter. During winter the importance of diffuse radiation appears to increase but the absolute value of total direct and diffuse SW↓ is relatively small at this time. There is at present no alternative to accepting this error, given the basic data available for the Chilton Valley during this study.

Both the actual and the adjusted values of shortwave radiation (Tables 2.1, 2.2 and Figure 2.2) may now be discussed. SW↓ values during the study year, for early spring, autumn and winter, are similar to the earlier data. However, November and January received higher amounts of radiation,

TABLE 2.1

MONTHLY MEAN VALUES OF SHORTWAVE RADIATION IN LY DAY^{-1} DURING THE STUDY PERIOD

Month	Maximum Possible at Recording Site	Radiation Received	% of Maximum Possible	Radiation Adjusted For Sky Line Obstruction 1969-70	Radiation Adjusted For Sky Line Obstruction 1965-67
August	296	203	68.8	243	*
September	429	239	55.8	257	283
October	558	418	75.0	438	375
November	684	516	75.1	543	413
December	730	480	65.8	507	538
January	696	508	73.0	535	468
February	586	437	74.8	459	462
March	424	269	63.3	284	336
April	252	216	85.7	259	225
May	135	119	88.0	198	182
June	74	55	74.9	138	128
July	115	83	71.8	137	138
August	211	137	65.0	196	*

* Data incomplete for August

TABLE 2.2

SEASONAL TOTALS OF SHORTWAVE RADIATION IN KILOLANGLEYS FOR THE STUDY PERIOD. FIGURES IN BRACKETS INDICATE PERCENTAGE OF THE TOTAL POSSIBLE

Summer 1969/70	Autumn 1970	Winter 1970	Spring 1969
42.86	18.53	6.42*	35.60
(71.3)	(74.0)	(-)	(72.8)

* Date are incomplete for August

TABLE 2.3

COMPUTATION OF DIRECT SHORTWAVE RADIATION LOST DUE TO THE PRESENCE OF HILLS. TIMES ARE LOCAL APPARENT TIMES

Date	Solar Declination (Degrees)	Times of Shading	Amount to be Subtracted due to shade (ly)	Total Incoming (ly)	% Lost
Dec 22	23.45	0000-0730 1900-2400	38.8	707.4	5.5
Jan 21, Nov 22	20.00	0000-0730 1830-2400	33.9	665.4	5.1
Feb 9, Nov 3	15.00	0000-0745 1815-2400	30.6	599.7	5.1
Feb 23, Oct 20	10.00	0000-0800 1815-2400	25.6	529.6	4.8
Mar 8, Oct 6	5.00	0000-0800 1730-2400	20.0	463.8	4.3
Mar 21, Sept 23	0.00	0000-0830 1700-2400	25.5	393.1	6.5
April 3, Sept 10	-5.00	0000-0830 1600-2400	25.2	327.7	7.8
April 16, Sept 10	-10.00	0000-0930 1530-2400	43.0	259.4	16.6
May 1, Aug 12	-15.00	0000-1000 1430-2400	53.1	196.7	27.1
May 21, July 24	-20.00	0000-1045 1345-2400	61.1	138.6	44.0
June 22	-23.45	0000-1130 1300-2400	71.5	104.3	68.2

TABLE 2.4

MONTLY DIRECT RADIATION LOST DUE TO SHADING

	Mean Daily Totals For Study Period ly day^{-1}	Percentage
August	40	16.6
September	18	7.1
October	20	4.5
November	28	5.1
December	27	5.5
January	27	5.1
February	23	5.0
March	15	5.4
April	43	16.6
May	79	40.0
June	83	60.0
July	55	40.0
August	59	30.0

while December received less than in the preceding years. The study year, therefore, appears atypical, when compared with the earlier years, in that it does not have its maximum energy input in December, but in November. The apparent possible variation of SW↓, and presumably cloud cover, in November, December and January, is important as it may have serious effects on other physical parameters such as soil moisture.

Daily shortwave radiation (Fig. 2.3) shows that there were some particularly low values recorded in mid-December, and seldom during this month were values persistently high. The daily data also can be used to explain some of the other anomalies in Fig. 2.2. A number of low values in September 1969, and some high values in November 1969 and in January 1970 are all reflected in the annual data. The most striking feature of the daily values however, is the lack of persistence in either high or low values. There is, therefore, a high variability in this important element of the climate.

The adjusted SW↓ for the Chilton Valley may be compared (Fig. 2.4) with data from other locations or obtained by other methods. SW↓ recorded in the Chilton Valley from February through to July, both in the study year and in the earlier years, is fairly similar to that recorded at Christchurch between 1960 and 1965, except for relatively low values at the Chilton Valley in March. However, from October to January the Christchurch values are higher than the Chilton Valley values, sometimes by over 100 $ly.day^{-1}$. This was not so during the study period except for the month of December. Since the data are for different periods a direct comparison

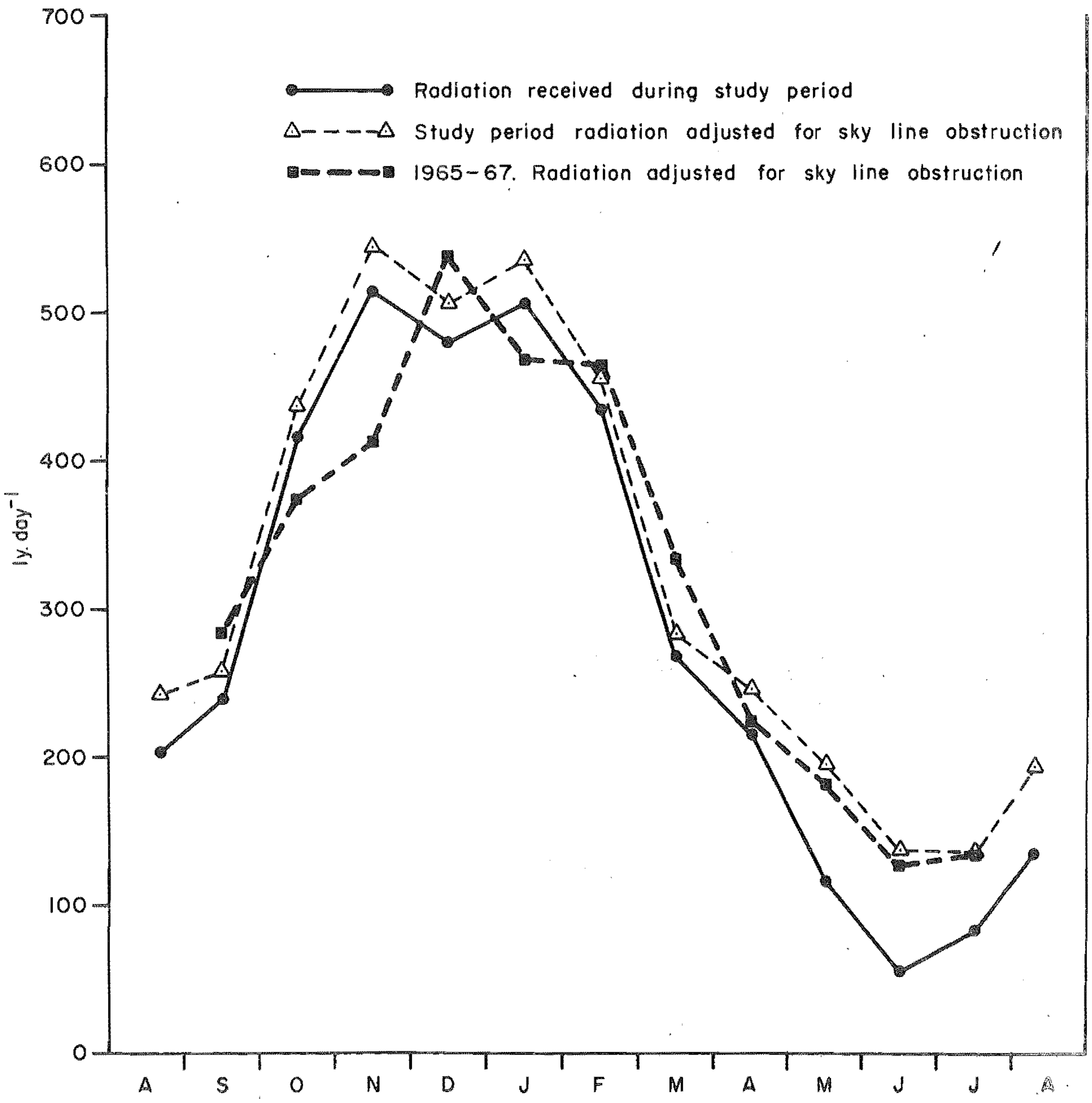


FIGURE 2.2: Monthly mean values of shortwave radiation received at the Chilton Valley

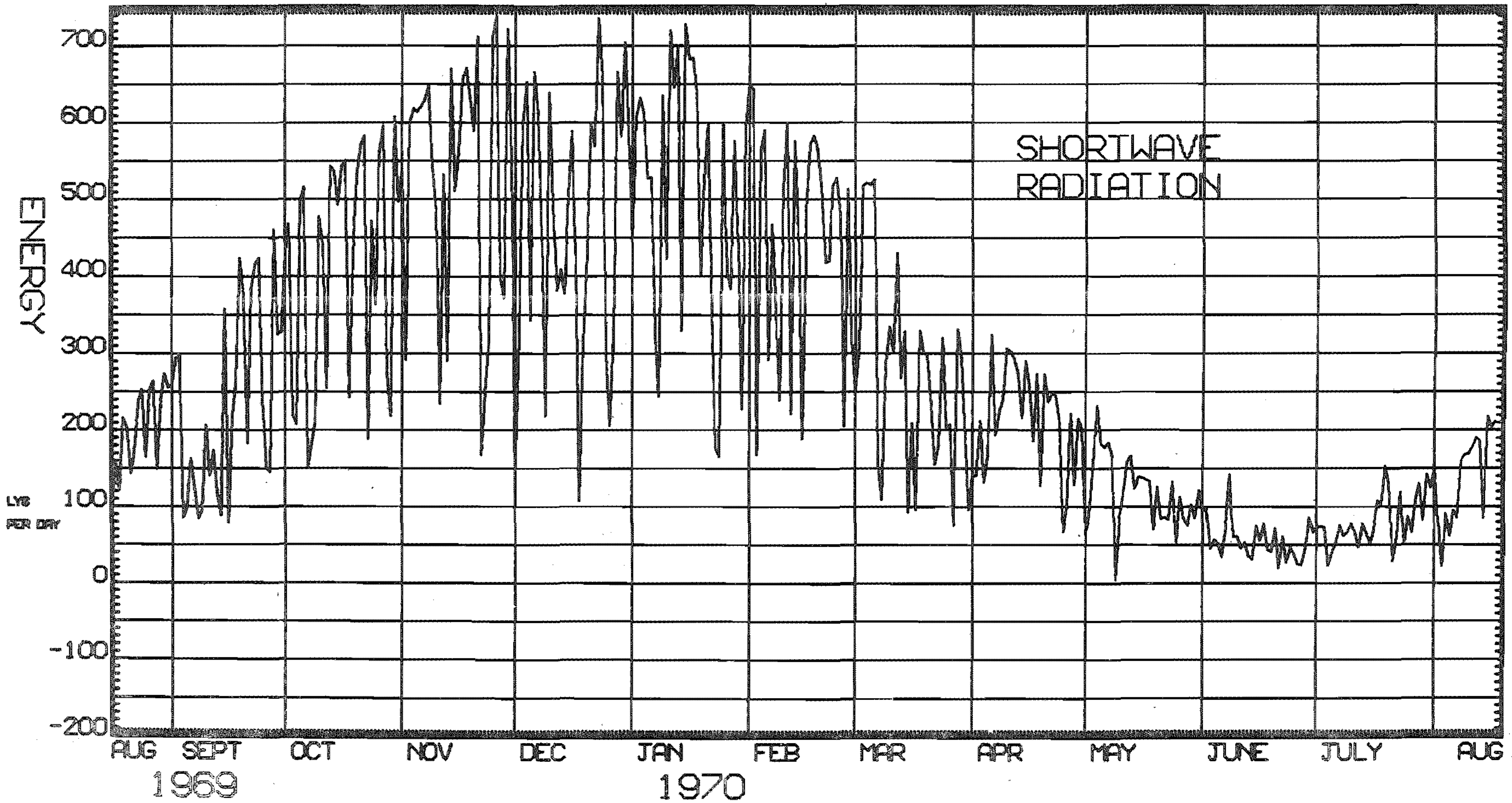


FIGURE 2.3: Daily values of shortwave radiation received at the Chilton Valley during the study period

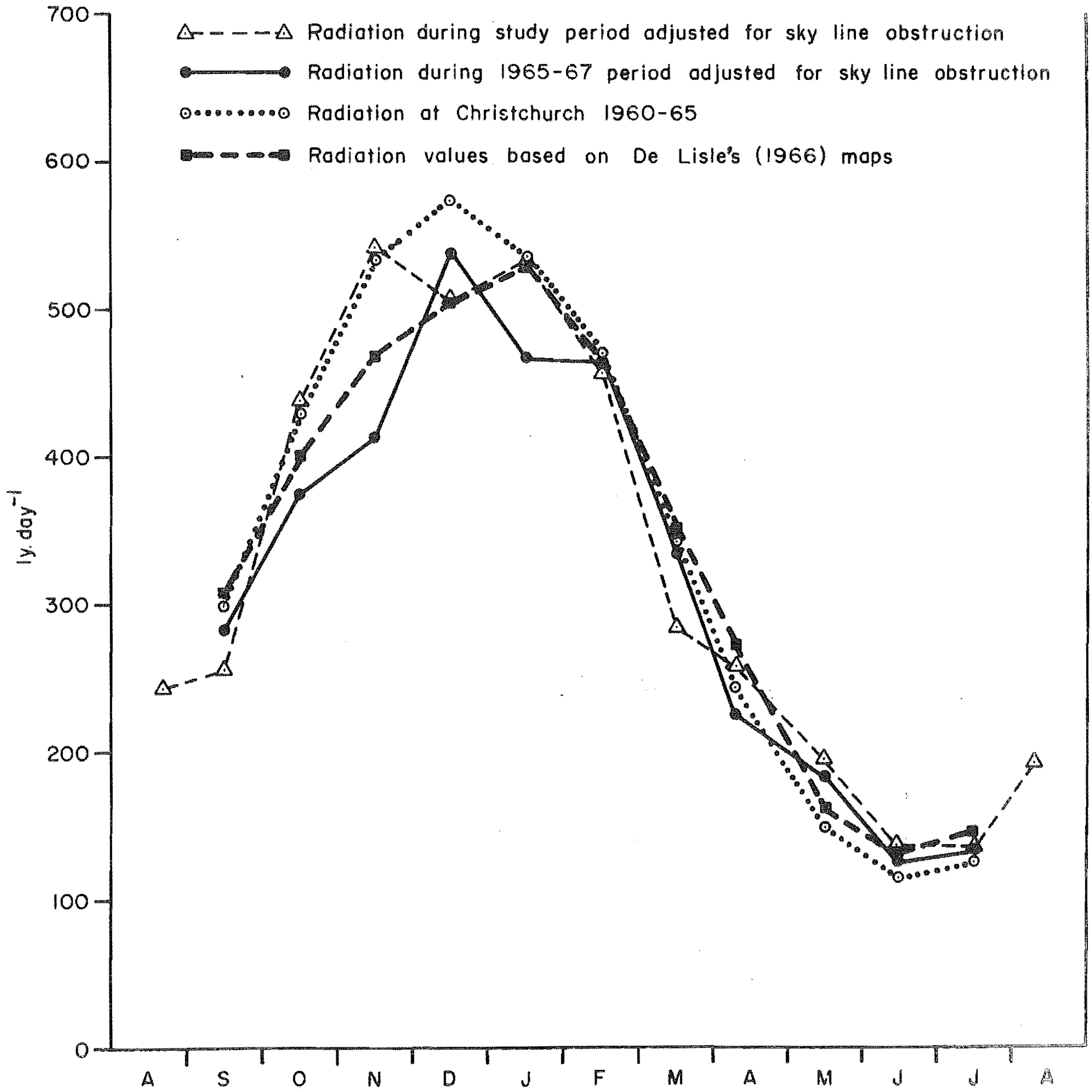


FIGURE 2.4: A comparison of shortwave radiation received at the Chilton Valley with that received at Christchurch Airport, and that received at the Chilton Valley according to de Lisle (1966).

cannot be made but these results suggest that, although for much of the year the Chilton Valley receives a similar amount of radiation to Christchurch, during the spring and a large part of the summer, cloud cover and/or increased atmospheric water vapour, give rise to lower radiation values in the alpine location. This suggestion is supported by work by de Lisle (1966), who has constructed monthly maps of shortwave radiation for New Zealand using a modified Angström formula and data in which over half the stations used had a sunshine record of between 20-25 years. Values for the Chilton Valley interpolated from these maps are also shown in Fig. 2.4. Once more, although there is similarity between Christchurch and the interpolated values for the Chilton Valley from February to July, the interpolated alpine values are lower for the spring and most of the summer. As mentioned in section 1.4, this is in agreement with the long period precipitation records that show maximum rainfall occurring in October. Earlier, in the comparison of the Chilton Valley 1965-67 data and the study year data, November, December and January of the latter appeared to have atypical values. The above comparison between Christchurch data, de Lisle's values, and the study year SW↓ at the Chilton Valley suggests, in particular, that the high value in November of the study year was unusual.

The relatively high spring and summer cloudiness and/or atmospheric water vapour content, apparent at the Chilton Valley site may also be expressed in terms of atmsopheric transmissivity. All other factors being equal, radiation amounts should increase with altitude due to increased values of atmospheric transmissivity. Geiger (1965), for example, reports that for the European Alps an increase of about 60 $ly.day^{-1}$, in summer, and 25 $ly.day^{-1}$, in winter, occurs over an altitude increase of 760 m. Although starting at a higher level, this increase in altitude is similar to that between Christchurch and the Chilton Valley. However, although a winter increase in SW↓ is noted at the valley, there appears to be generally less SW↓ in the summer. Therefore, at the Chilton Valley in summer, SW↓ appears to be affected more by an increased atmospheric water content than by the decreased atmospheric mass associated with the altitude of the site.

At two points in this chapter, values of clear sky atmospheric transmissivity, T, are assumed. A value of 0.75 is taken for the adjustment for sky line obstruction computation, and a value of 0.80 is selected in calculating daily values of Q for different slopes (section 2.9). The difference in T values reflects the fact that in the horizon corrections, times of high solar zenith angle, and therefore low T, are important, but in the computation of daily totals, a higher average zenith angle would lead to increased T values. Both values selected are compatible with a range of 0.72 to 0.81 found for Wellington near sea level (de Lisle,1966). Computed T values for the Chilton Valley on 25 November, 22 December 1969 and 14 January 1970 are 0.78, 0.78 and 0.77 respectively. This computation, based on a solar constant of 2.0 $ly.min^{-1}$, uses adjusted SW↓ values for the Chilton Valley and therefore includes an assumed T value in allowance for sky line obstruction. However, near the time of the summer solstice, the horizon adjustment involves less than 6% of the total SW↓ (Table 2.3). In view of this, the above computed T values will not suffer severely from the previous incorporation of an assumed value of T. The computed T values are seen to be within 3% of the values selected for sky line adjustment and Q reception on a slope calculation. The values of T selected for these calculations therefore appear reasonable.

Finally in this section, it is of interest to examine the relative proportion of direct and diffuse radiation received in the Chilton Valley. In the absence of measurements using a shade ring on the radiometer, the q/Q ratio may be approximated by a method suggested by Rayner (Soons and Rayner, 1968). Shortwave radiation recordings immediately before and after the effective sunrise are assumed to be representative of diffuse radiation, and direct plus diffuse incoming shortwave radiation, respectively. Using this assumption, the proportion of diffuse to direct radiation can be calculated. An attempt was made to compute the q/Q ratio on or near the dates of solar declinations used in the adjustment for horizon calculations (Table 2.3). This was not always possible due to the absence of cloudless conditions at times of effective sunrise. The results for the times when the computation of the q/Q ratio was possible are shown in Fig. 2.5. The value of q/Q is about 10% for all seasons except winter when it rose to over 20%. This is similar to values computed from measurements of q and Q for clear days made by Funk (1963) for Aspendale, Australia. Using four years data he found a variation of q/Q from 12% in November to 19% in May, although the June value fell to 13%.

It should be noted that the proportion of diffuse

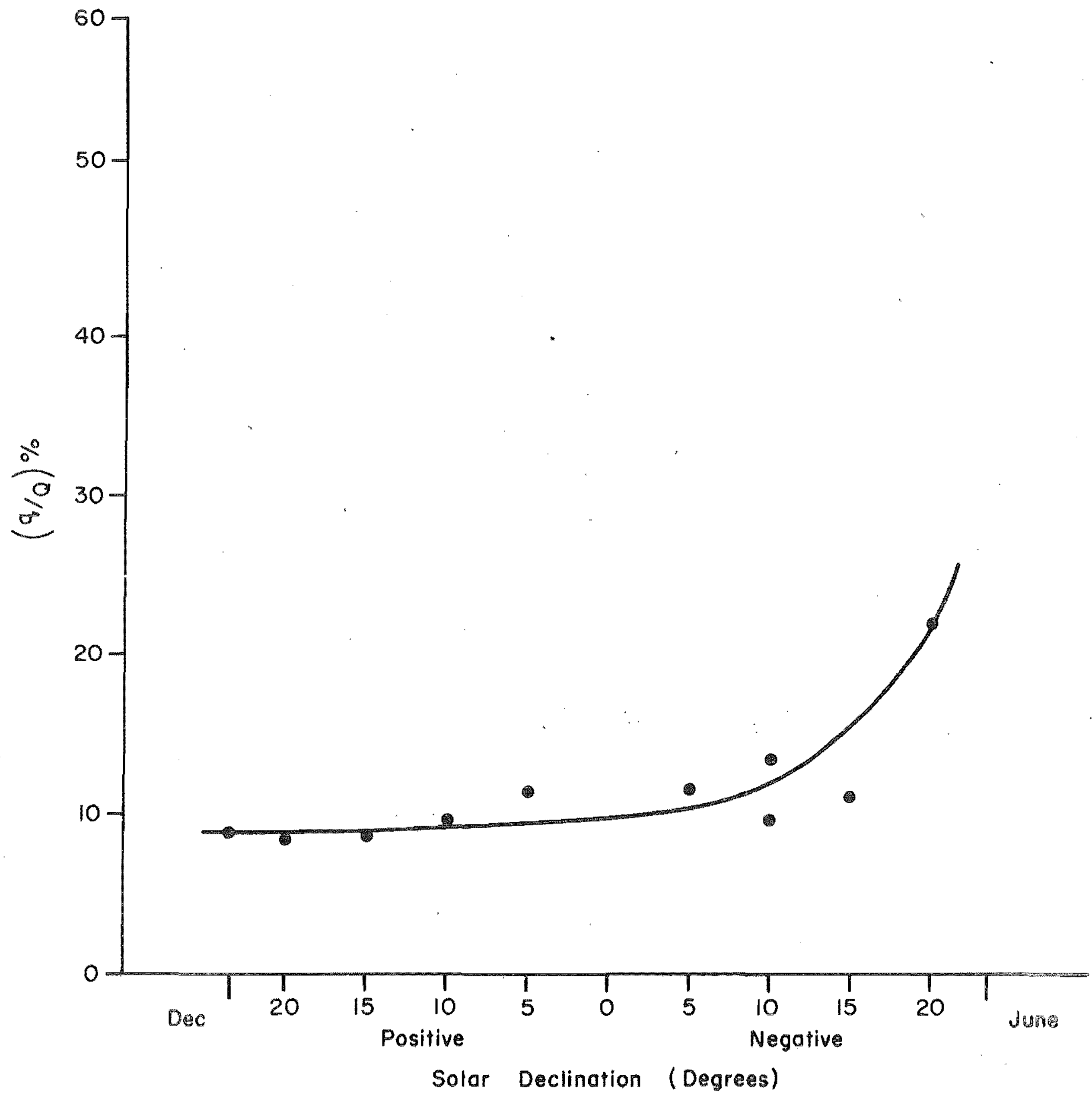


FIGURE 2.5: Values of (q/Q) in percent at the Chilton Valley for cloudless days at different values of solar declination during the study period

radiation is usually much higher on cloudy days, as has been documented by Liu and Jordan (1960). Jackson (1967) points out that at any season at Taita (North Island, New Zealand) diffuse radiation may contribute 50% of the total radiation received on some slopes, and is the only source of solar radiation on steep south facing slopes in mid winter. It is possible that similar conditions apply to the Chilton Valley (see section 2.9).

2.3 Albedo

Albedo measurements were made on four individual days in four different seasons. These were made on clear days so as to obtain maximum deflection on the recorder. Tests were made over different vegetation and surface cover types and an attempt was made to see if the height of the albedometer above the ground affected the α values obtained. In all cases, except where noted, the albedometer was fixed at about 80 cm above the ground. Owing to the inhomogeneity and the slope of the surface, the descriptive titles refer only to the dominant feature of the surface sampled. The surface of the valley bottom as a whole is composed mainly of tussock grass. Thus the albedo for this type of vegetation is probably the most representative, and is used in the net radiation models in section 2.5. Finally, a continuous daily record was taken over one surface, in order to examine the diurnal variation in α.

Albedos were measured over different surfaces on January 29th 1967, 9th April 1970, 16th July 1969, and 4th October 1969 (Table 2.5), and are assumed to be representative of summer, autumn, winter and spring respectively. All of the

TABLE 2.5

ALBEDO OF DIFFERENT SURFACES
(PERCENT)

Surface	Oct 4 1969 Spring	Jan 29 1967 Summer	April 9 1970 Autumn	July 16 1969 Winter
Tussock	14.0	12.7	14.6	65.0
Wet Tussock		12.3		
Matagouri		13.3	13.7	61.0
Rocks and lichen	15.2	16.1	15.2	61.7
Cassinia	14.2	17.6	12.4	66.6
Snow Totara		13.7	12.6	
Brown Earth	14.6	10.2	14.7	
Wet Brown Earth		15.0	12.9	58.0
Grey stones	14.6	18.6	18.9	
Wet Grey stones		10.7	14.8	61.0
Old snow from 20 cms				65.0

measurements were made at about solar noon. Repetition of measurements for one surface indicated that the standard deviation of such measurements is 0.3%.

There is little change in albedo of most surfaces between the dates in the summer, autumn and spring seasons. The variation for tussock, for example, is less than 2% and that for rocks and lichen is less than 1%. The variation occurring when some surfaces were artificially wetted was often greater than the albedo difference of the same surface at the three different times of the year. The surface of grey stones is a case in point. The variation between different types of surface cover within the valley had a range of about 10% on January 29 1967 and less on the other dates. This is quite a common order of range of albedos for natural surfaces (see for example, Sellers, 1965). The absolute values of the albedos taken on the summer date are 10.2% to 18.6% with an average of 14.6% and a value for tussock, the most widespread surface, of 12.7%. These values are lower than those commonly found for agricultural crops (Davies and Buttimore, (1969) suggest a value of 25%) and also lower than the 18.0 to 20.0% recorded for alpine tundra at 3580 m (Terjung et.al. 1969). The present values are more in line with the values of 14.6 and 14.8% found for relatively low altitude forest and grassland between latitudes 45^{o} and 55^{o} N in North America (Kung et.al., 1964). The albedos for the Chilton Valley surfaces may prove typical for much of the South Island High Country but more measurements are needed. In particular, the albedos of the surfaces of forests and lakes have not been measured.

Possibly the most striking feature emerging from Table 2.5 is the high albedo values on the winter date. These values were recorded at a time when patches of old snow were lying on the ground, and it is estimated that 40-60% of the surface area of the valley bottom was snow covered. The rest of the surface was thoroughly saturated by snow meltwater. The albedos for the different kinds of surface cover, indicated in the table, were taken in positions identical, wherever possible, to those of the January, April and October data. The sun angle (90° - zenith) for January was 67°, while that for July was 26°. The readings were taken on a slope of 14° and the true angles of incidence for January and July were 53° and 12° respectively. The angles of incidence and the snow cover easily explain the high albedos, which average 62.6% for all of the surfaces tested. This value of albedo falls well within the range for old snow (40 - 70%; Geiger,1965). It is reasonable to assume that albedos in the Chilton Valley can pass through an even larger range representing those of fresh snow, old, and melting snow surfaces. During the study period, it is estimated that snow lay on the ground for ten days (section 2.4). The relatively high albedos assumed to have existed during this time have been taken into account in obtaining the net radiation for this period.

One more feature from Table 2.5 is worth mentioning briefly. The effect of moisture can decrease the albedo of a non-vegetated surface by several percent as can be seen in the case of grey stones on January 29, and brown earth on April 9. However, if the soil surface is completely saturated, and more water is applied, a shiny surface can be produced, and this can result in an increased albedo, as in the case of wet brown earth for the January date. A similar shiny surface for the brown earth areas was observed to exist due to snow melt water on July 16 when the winter observations were made.

An examination was made to see if there was a variation of albedo on a diurnal time scale. Albedo measurements for the tussock and short grass surface under the net radiometer stand at the main site (Fig. 1.4), were made at 7½ minute intervals throughout the day of 18 November 1969. A diurnal variation in albedo was apparent, as can be seen in Table 2.6. Such a variation, although not necessarily of the same magnitude, has been found by many other workers (Davies,1967a; Wendler and Streten 1969, Terjung et.al.,1969). The implication of the existence of such a diurnal variation of albedo for one surface, is that daily average values may be higher than the values recorded near solar noon (Table 2.5).

Where the different surface covers occupy only a small area, the influence of other types of surface near to that which is being measured may have an effect on the value of albedo obtained. In order to investigate what effect reflection from nearby surfaces did, in fact, have, measurements at different heights were taken over a tussock dominated surface at the recording site on 4 October 1969. Taking albedo records at increasing heights had the effect of increasingly integrating radiation from other surfaces with that of the principal surface. The results (Table 2.7) show a slight increase of albedo with height, but the difference in albedo, 1.3%, is small. This adds a degree of confidence

TABLE 2.6

VARIATION OF ALBEDO WITH SOLAR ZENITH ANGLE FOR TUSSOCK AND SHORT GRASS SURFACE AT THE MAIN RECORDING SITE

	Solar Zenith Angle (Degrees)					
	30	40	50	60	70	80
Albedo %	16	17	19	21	25	50

TABLE 2.7

VARIATION OF ALBEDO WITH HEIGHT OF MEASUREMENT ABOVE A TUSSOCK AND SHORT GRASS SURFACE AT THE MAIN RECORDING SITE

Height m	Albedo %
0.71	12.4
1.52	12.4
3.05	13.7

to the records taken at 80 cm.

2.4 Net Shortwave Radiation

Since net shortwave radiation is used as the forcing function in Lettau's climatonomy (section 6.3) it is appropriate to derive its values at this stage. Following the above discussion of albedo values, and bearing in mind that tussock is the most common surface, it was decided to take an albedo value of 0.13 for all of the year, except for times of snow cover. It was assumed that when precipitation occurred, and the daily mean air temperature in the screen was less than 0° C, snow had fallen. It was further assumed that some snow cover lasted on the ground for two days after the daily mean air temperature had risen above 0° C. Such conditions were met only three times during the study period; 21-23 August 1969, 10-12 October 1969 and 15-18 July 1970. On the latter period, observation confirmed the existence of snow lying on the ground. During these periods an albedo of 0.50 was assumed. The net shortwave radiation was computed as being $(Q+q)(1-\alpha)$.

Daily values of net shortwave radiation (Fig. 2.6) and the monthly means (Fig. 2.18 and Table 2.9), are directly proportional to values of SW↓ in almost every case, and therefore reflect the changes found in the latter data. The ten days when the constant of proportionality $(1-\alpha)$ differs, occur in three individual months, and the different value of the albedo for these days has little effect on the monthly means for these months.

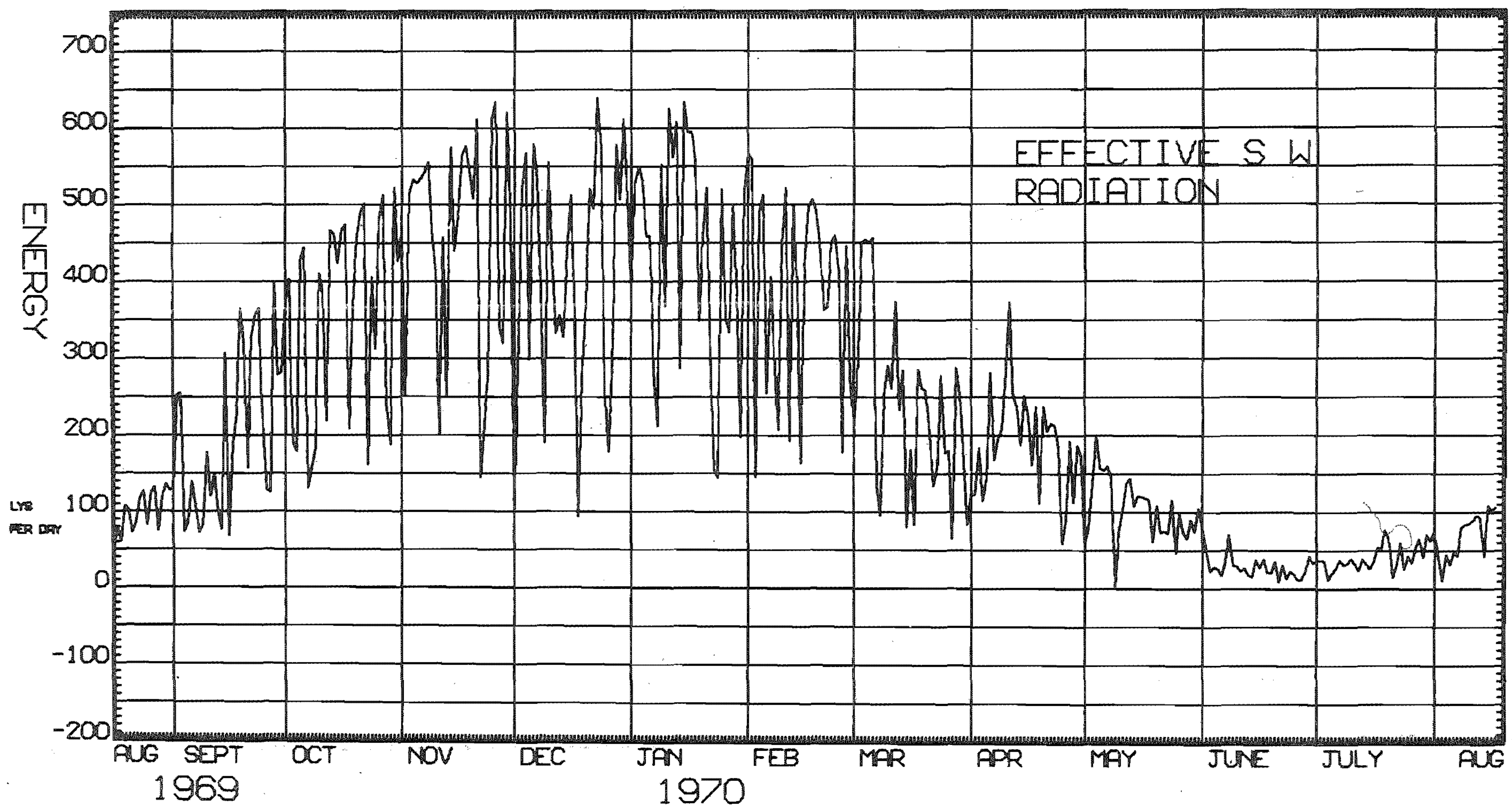


FIGURE 2.6: Daily values of net shortwave radiation at the Chilton Valley during the study period

2.5 The Relation Between Shortwave Radiation and Net Radiation

In order to complete the radiation balance equation data are required on the longwave fluxes. It was, however, impracticable to measure these on a continuous basis. Indeed, under the conditions in the Chilton Valley it was not possible to measure continuously the net radiation itself (see section 1.5). Therefore daily values of net radiation were acquired by means of regression relationships between net radiation and incoming shortwave radiation. Before describing the particular method that is used for the Chilton Valley data, it is important that the nature and limitations of radiation regression relationships be considered.

Three types of relationship are available. The simplest is a linear regression of the form

$$Rn = A\ (Q + q) - B \qquad \text{\_\_\_\_ 2.5.1}$$

where A and B are the regression coefficients. This has been used by Davies in several studies (1967a, 1967b) (Davies and Bultimore, 1969). Surface albedo has also been included by Montieth and Szeicz (1961) as follows:

$$Rn = a\ (1 - \alpha)\ (Q + q) + b \qquad \text{\_\_\_\_ 2.5.2.}$$

where the lower case letters denote possibly different regression coefficients. Gay (1969) has shown that in the above equations the dependent variable, Rn, also contains the independent variable (Q + q), and thus high correlations are to be expected. While not eliminating the problem, Gay introduces a further type of equation:-

$$Rn = (1 + \lambda)\ (1 - \alpha)\ (Q + q) + b \qquad \text{\_\_\_\_ 2.5.3.}$$

where λ is a longwave exchange coefficient numerically equal to A $(1 - \alpha)^{-1} - 1$. Besides Gay's criticism of relationships between shortwave and net radiation, Idso et.al. (1969) have drawn attention to the fact that comparisons of the regression coefficients in equations such as 2.5.1. are only reliable if there are a significant number of data points near the origin. Furthermore, Hay (1970) has demonstrated a marked hysteresis effect in the relationship between longwave radiation and shortwave radiation over the period of a year.

In the present study attention is given to the establishment of a relationship for obtaining daily totals of Rn from Q + q for the Chilton Valley site. A series of 145 days of net radiation and shortwave radiation totals was available from 1964 and 1965 when the net radiometer was working eficiently, although not attended daily, and before it was damaged by opossums for the first time. These daily totals were subjected to regression analyses using equations of the form of the equations 2.5.1. and 2.5.2.

A comparison of the equations of type 2.5.1. and 2.5.2. (Table 2.8 and Figs. 2.7 - 2.16) shows that, in general, there is no difference in intercept values, correlation coefficients (C.C), or standard errors of estimate (S.E.E) except in the case of equations containing data from all seasons (Fig. 2.7 and 2.12 and equations 2.5.4. and 2.5.9.). Where a value of α of 0.13 has been taken for all days of data, only the gradient of the two types of equations is altered. These results can be predicted algebraically if α is held constant. In other work where α is measured daily (e.g. Davies and Bultimore,1969), the absence of any large variation in α values

TABLE 2.8

NET RADIATION AND SHORTWAVE RADIATION REGRESSIONS AND RELATED PARAMETERS FOR CHILTON VALLEY DATA OF 1964-65. (Rn AND SW↓ IN LY DAY^{-1} IN ALL CASES)

Equation Number	Figure Number	Regression Equation	Number of Observations	C.C.	S.E.E. (ly day^{-1})	Applicability
2.5.4	2.7	Rn = 0.66 SW↓ -68.2	145	0.96	38.1	All seasons
2.5.5	2.8	Rn = 0.52 SW↓ -13.8	75	0.92	31.7	Spring (Sept. - Nov.)
2.5.6	2.9	Rn = 0.57 SW↓ -9.1	51	0.96	24.1	Summer (Dec - Feb)
2.5.7	2.10	Rn = 0.88 SW↓ -120.0	19	0.38	34.4	Winter (June - July)
2.5.8	2.11	Rn = 0.57 SW↓ -22.8	126	0.94	31.9	Spring and Summer
2.5.9	2.12	Rn = 0.74(1-α) SW↓-60.7	145	0.96	36.3	All seasons
2.5.10	2.13	Rn = 0.60(1-α) SW↓-13.8	75	0.92	31.7	Spring (Sept - Nov)
2.5.11	2.14	Rn = 0.66(1-α) SW↓-9.1	51	0.96	24.1	Summer (Dec - Feb)
2.5.12	2.15	Rn = 1.02(1-α) SW↓-120.0	19	0.38	34.4	Winter (June - July)
2.5.13	2.16	Rn = 0.65(1-α) SW↓-22.8	126	0.94	31.9	Spring and Summer

FIG. 2.7

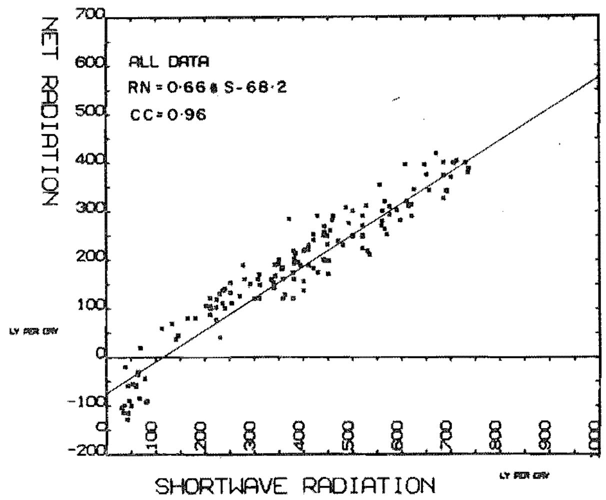
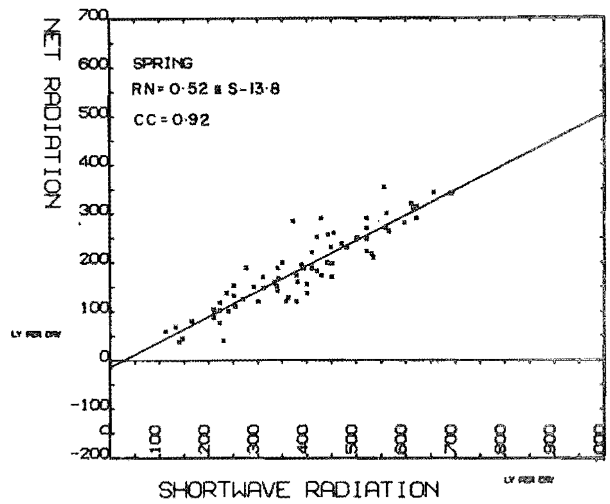


FIG. 2.8



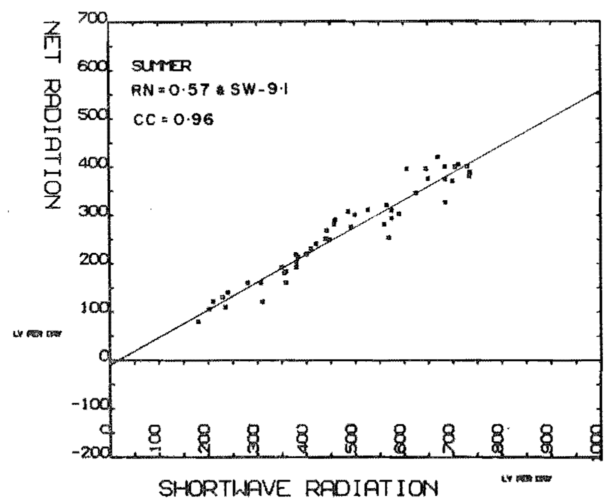


FIG. 2.9

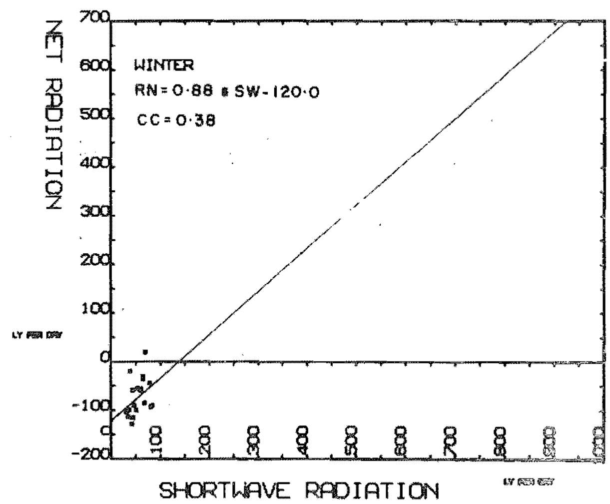


FIG. 2.10

FIGURES 2.7 - 2.10: Net radiation/Solar radiation relationships for all data (Fig. 2.7), the Spring season (Fig. 2.8), the Summer season (Fig. 2.9), and the Winter season (Fig. 2.10).

FIG. 2.11

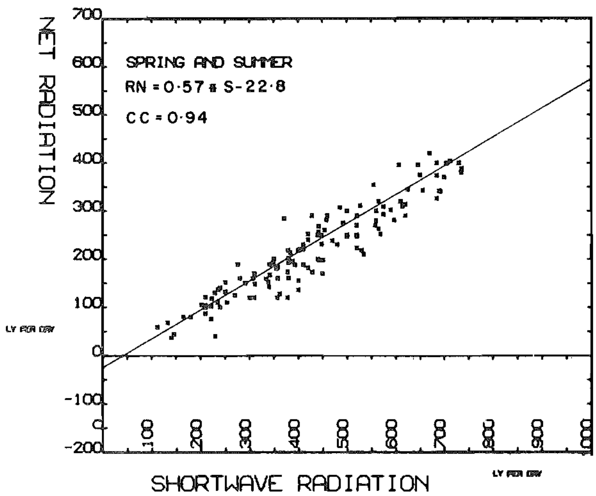
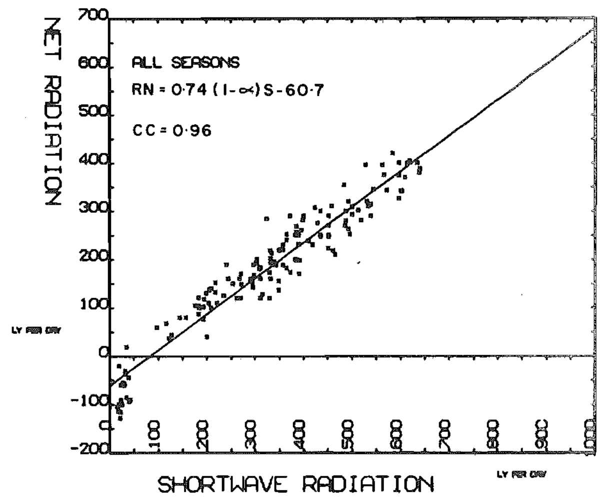


FIG. 2.12



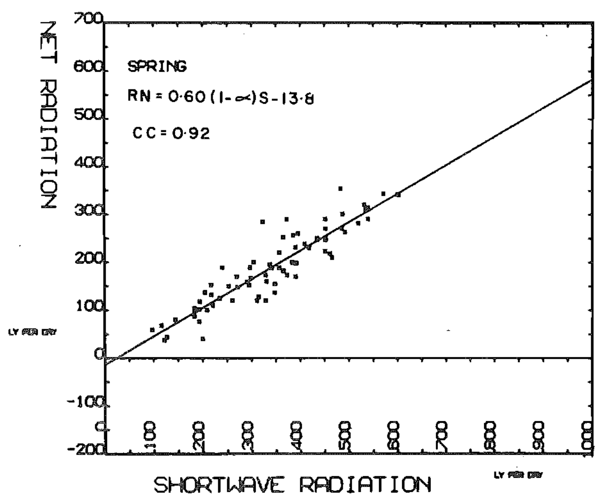


FIG. 2.13

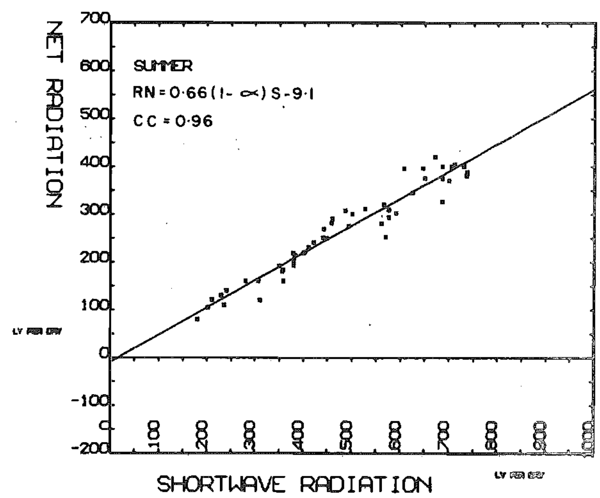


FIG. 2.14

FIGURES 2.11 - 2.14: Net radiation/Solar radiation relationships for the Spring and Summer seasons (Fig. 2.11), all seasons (Fig. 2.12), the Spring season (Fig. 2.13), and the Summer season (Fig. 2.14).

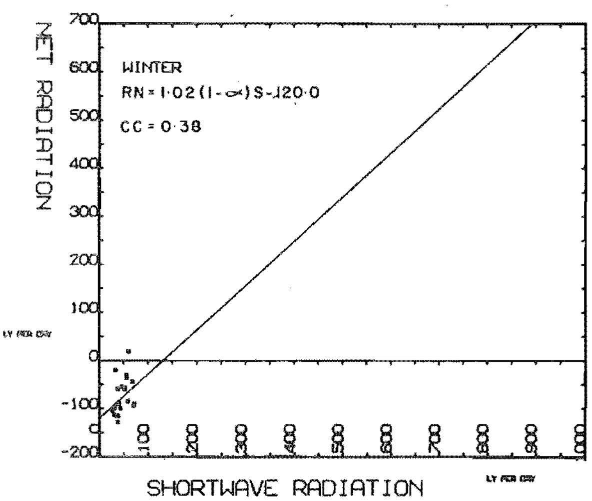


FIG. 2.15

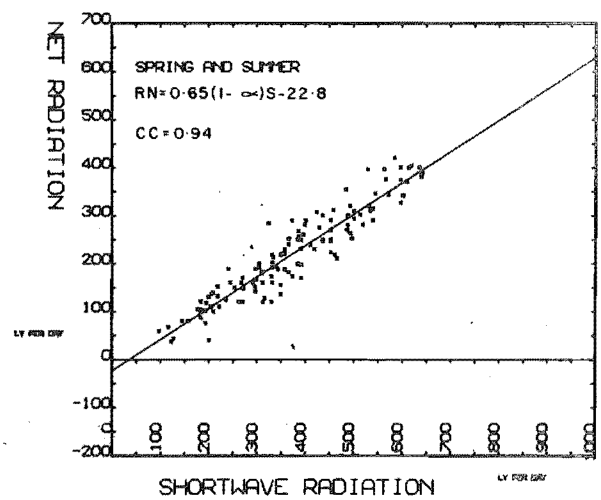


FIG. 2.16

FIGURES 2.15 - 2.16: Net radiation/Solar radiation relationships for the Winter season (Fig. 2.15), and the Spring and Summer seasons (Fig. 2.16).

leads to a similar lack of benefit from the inclusion of the albedo term. However, if α does show a large day to day variation, advantage can be gained by its inclusion in the regression equation. This is shown by a comparison of equations 2.5.4. and 2.5.9. In obtaining equation 2.5.4. α was not included but equation 2.5.9. results when α is incorporated with values of 0.13 for spring and summer data and 0.50 for winter data. It can be seen that besides the gradient change, intercept values show a difference of 7.5 ly day^{-1}. Although there is no change in the correlation coefficients of the two regressions, the inclusion of the albedo term gives a small improvement in the standard error of estimate.

The winter data points used in the analysis appear to lower the intercept value in the regression. This occurs irrespective of whether or not albedo is included in the regression analysis. The winter Rn values (Fig. 2.10) are very low, and when all the data are combined (e.g. Fig. 2.7, equation 2.5.4.), the winter data produce a larger negative value for the intercept, than when they are excluded from the analysis (Fig. 2.11 and equation 2.5.8.). The winter values alone (Fig. 2.10) produce a very low intercept value (-120.0 ly day^{-1}). Other workers (e.g. Hay, 1970) have not found such large negative intercept values. Those found in the present study could be real or else due to observational error. In support of the latter it is noted that frequent frost on the radiometers would produce such an effect. There is therefore some doubt as to the validity of the winter data, owing to their profound effect on the nature of the regression equations formed from them. Accordingly attention is given to relationships using the spring and summer data only.

Having rejected the winter data it is possible to use either the spring and summer regressions (equations 2.5.5. and 2.5.6.) separately, or to use the combined spring and summer data regression (equation 2.5.8.) in order to obtain Rn values for the study period. When the two separate regressions were used, with the spring regression being applied to the winter season, and the summer regression being applied to the autumn season, a discontinuity was apparent, in values of Rn, at the changeover point between the application of the different regressions. This gave rise, for example, to November having only the third highest monthly Rn total, whereas it received the highest total of SW↓ of all months. In order to prevent such discontinuities, it was decided to use the combined spring and summer regression equation, despite the fact that the S.E.E. of the latter is slightly higher (Table 2.8).

Since it has been shown above that albedo can be important at times of snow cover, an attempt to accommodate this influence was made. The actual equation used to obtain daily Rn rotals from shortwave radiation totals was equation 2.5.13. which includes the albedo term. An albedo value of 0.13 was used for all days except those when it was estimated, using the method described in the previous section, that snow was lying on the ground. On these days an albedo of 0.50 was used.

The choice of equation 2.5.13. has been determined by the data available, and the need to incorporate the effect of albedo on days when snow lay on the ground. With regard to the work of other investigators of Rn/SW↓ relationships, the following points are relevant. The criticism of Idso et.al. (1969) is not fully met as Fig. 2.16 shows that there is not an equal distribution of data points extending to the origin. The use of an alternative method suggested by Fritschen (1967), of employing pooled hourly regressions, was prohibited by lack of access to hourly data for the period from which the present regression equations were determined. Since, for the majority of days, the albedo is considered constant, the use of Gay's relation, equation 2.5.3., would not add any accuracy, and again only the gradient of the regression equations would be altered. In the present study the deficiencies of the Rn/SW↓ linear regression relationship are recognised, but a lack of other alternatives demands its use. However, the relation employed (equation 2.5.13.) is based on data collected from the location for which it is used, and under such conditions, several workers (e.g. Davies and Buttimore, 1969) believe that the Rn/SW↓ relation is of value for obtaining daily totals of Rn.

2.6 Net Radiation During the Study Period

Daily totals of net radiation for the study period were computed using the method described in the previous section. Both daily totals (Fig. 2.17) and monthly means (Fig. 2.18 and Table 2.9) follow, in almost all cases, the trends shown by SW↓ . This is to be expected because the same values for the regression constants and the albedo (except for days of lying snow) were applied throughout the year, and also

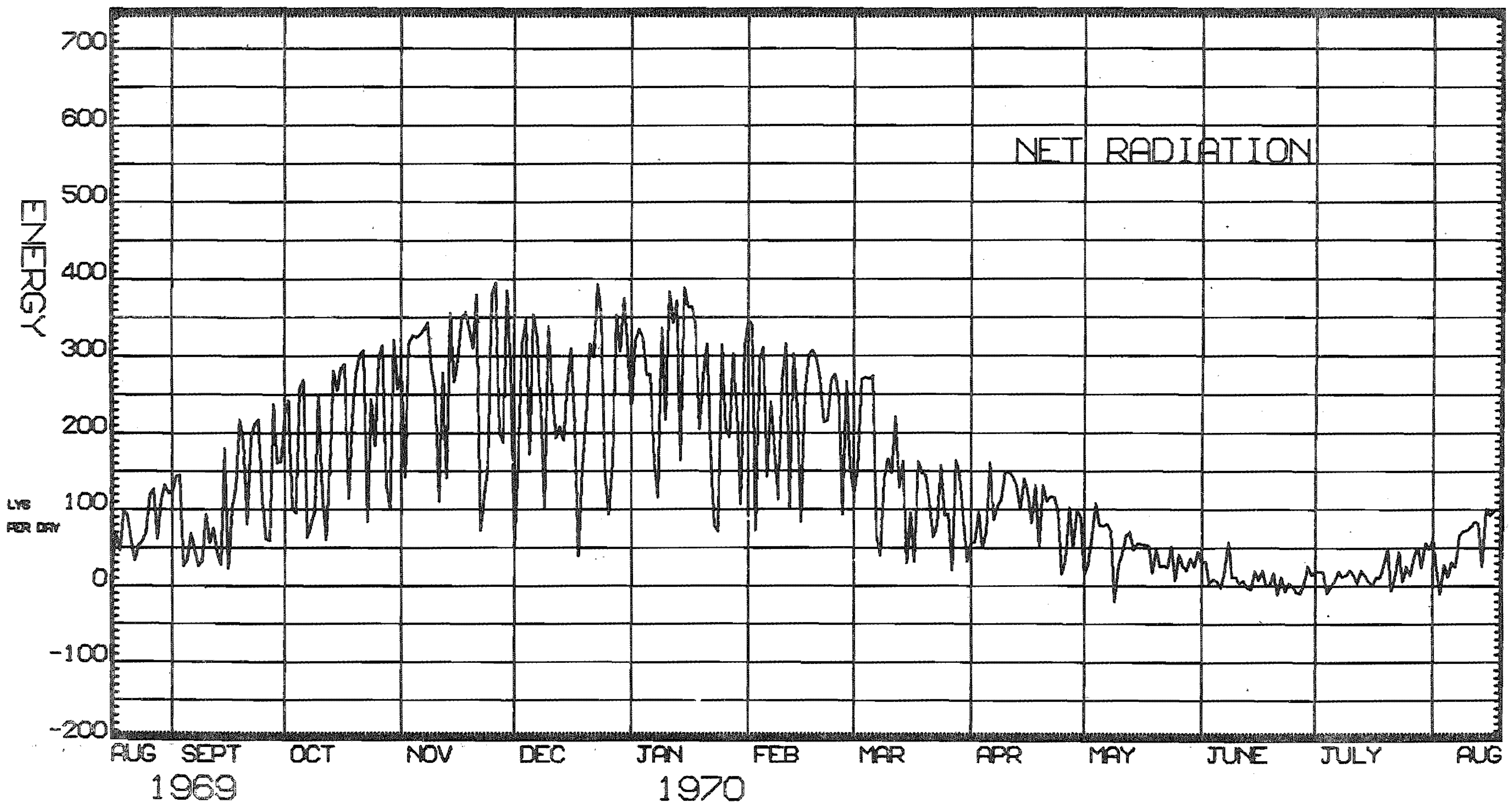


FIGURE 2.17: Daily values of net radiation at the Chilton Valley during the study period

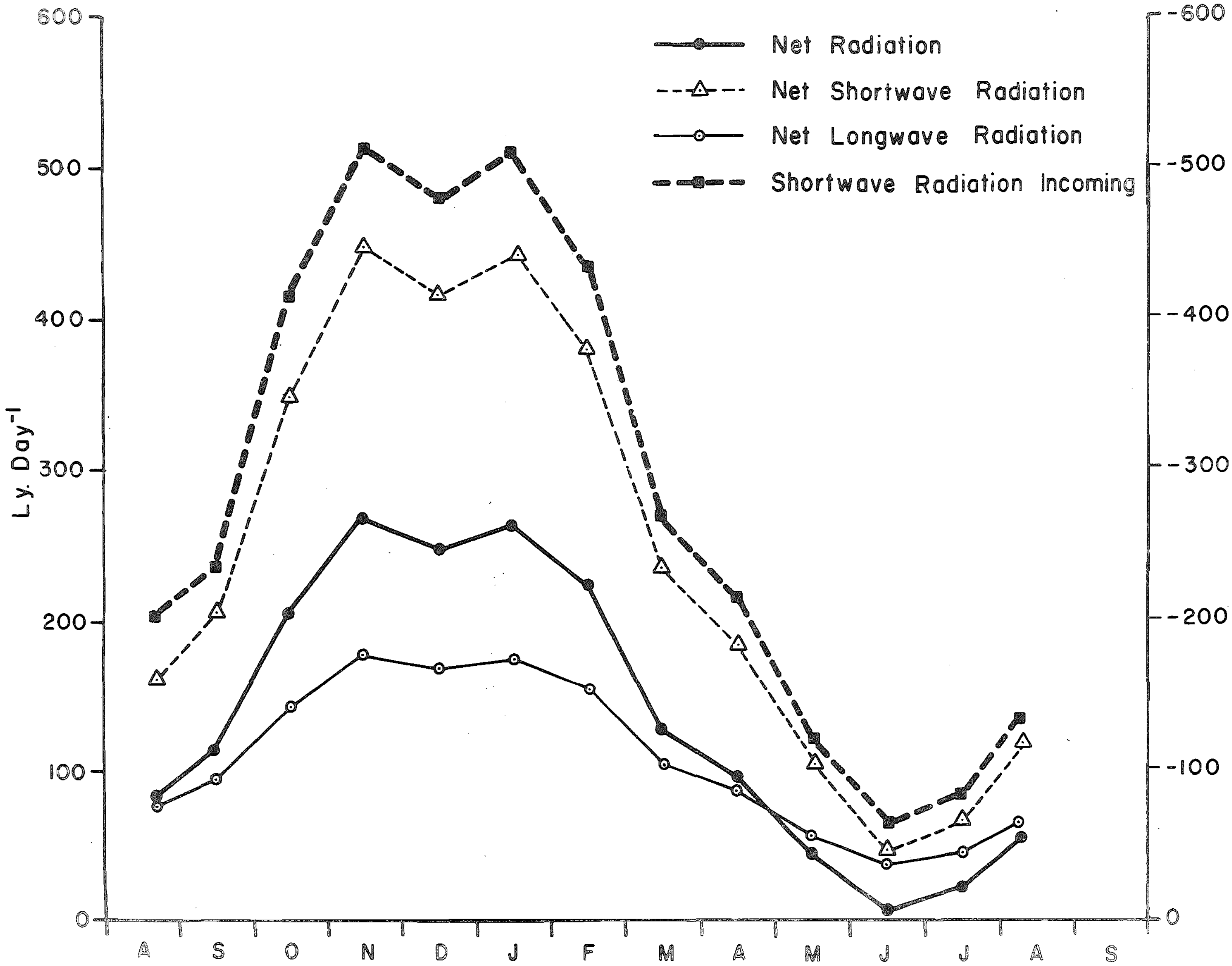


FIGURE 2.18: Monthly mean radiation values at the Chilton Valley during the study period. Right hand scale refers to net longwave radiation

because the individual components of Rn such as SW↓ and SW↑ are normally strongly correlated with it. The general observations on annual and daily variability of SW↓ (section 2.2) also, therefore, apply to the Rn values.

Mean monthly values of Rn for the Chilton Valley in the study year vary from 9 ly day^{-1} in June to 269 ly day^{-1} in November. These values are in the same order as those computed by Jackson (1967) for the Taita catchment (latitude 41° 11'S) for the year of 1966. He found a variation in mean monthly Rn of between approximately -5 to 290 ly day^{-1} in June and January respectively, if an albedo of 0.10 was assumed. It is interesting that despite the occurrence of a negative radiation balance on several days during the winter (see Fig. 2.17), the mean daily Rn for June is positive. Similar results are reported by Samukashvili (1969a) who worked at an altitude of 550m and at latitude 43° 29'N. Comparison of Figs. 2.3 and 2.17 shows that during the summer months the difference between SW↓ and Rn in the Chilton Valley, can be more than 300 ly day^{-1}. This may be due to the high surface temperatures that can occur (see Section 3.4) and the consequently large values of LW↑. Also, on clear days, a lack of clouds may be accompanied by relatively low LW↓ values.

Care was taken in choosing the method of computing the Rn data described above, since the Rn values are fundamental to this study. It is worthwhile emphasising at this point, that in radiation and energy balance climatologies Rn is the most important term. As Szeicz (1968) has pointed out, Rn represents the energy available for evaporation, melting snow and heating the soil and air.

TABLE 2.9

MONTHLY MEAN RADIATION VALUES IN LY DAY^{-1}

	Shortwave Radiation	Net Radiation	Net Shortwave Radiation	Net Longwave Radiation
August	203	83	163	-80
September	239	112	208	-96
October	418	204	349	-145
November	516	269	449	-180
December	480	249	418	-169
January	508	265	442	-177
February	437	224	380	-156
March	269	129	234	-105
April	216	97	184	-87
May	119	45	104	-59
June	55	9	48	-39
July	83	21	67	-46
August	137	55	120	-65

2.7 Net Longwave Radiation

Having computed values of net radiation and net shortwave radiation it is possible, with the use of equation 2.1.1., to calculate values of net longwave radiation (Fig. 2.18, 2.19 and Table 2.9). Considerable day to day variation is again found. However, by comparing Figs. 2.19, 2.17 and 2.6, it is seen that daily variation in net longwave radiation, in terms of absolute values, is less than that in either net shortwave radiation or net radiation.

Monthly mean values of net longwave radiation for the Chilton Valley in the study year range from -39 ly day^{-1} in July, to -180 ly day^{-1} in November. Jackson (1967) for the Taita catchment reported a variation of between -75 ly day^{-1} in August, to -125 ly day^{-1} in March and October 1966. The July value for the Chilton Valley appears rather low but a study by Lauscher (1934) suggests that this may, in part, be due to the effect of sky line obstruction. Although the same effect occurs in the summer, the November value for the Chilton Valley is in line with summer values (for clear sky conditions) that can be found in the highland areas of the western United States (Sellers, 1965 p.60).

2.8 The Relative Sizes of the Components of the Radiation Balance

The data available on many of the components of the radiation balance makes it possible to examine the relative importance of these components in terms of magnitude. The first reason for doing this, in the present study, is to emphasise that the important net radiation term is itself

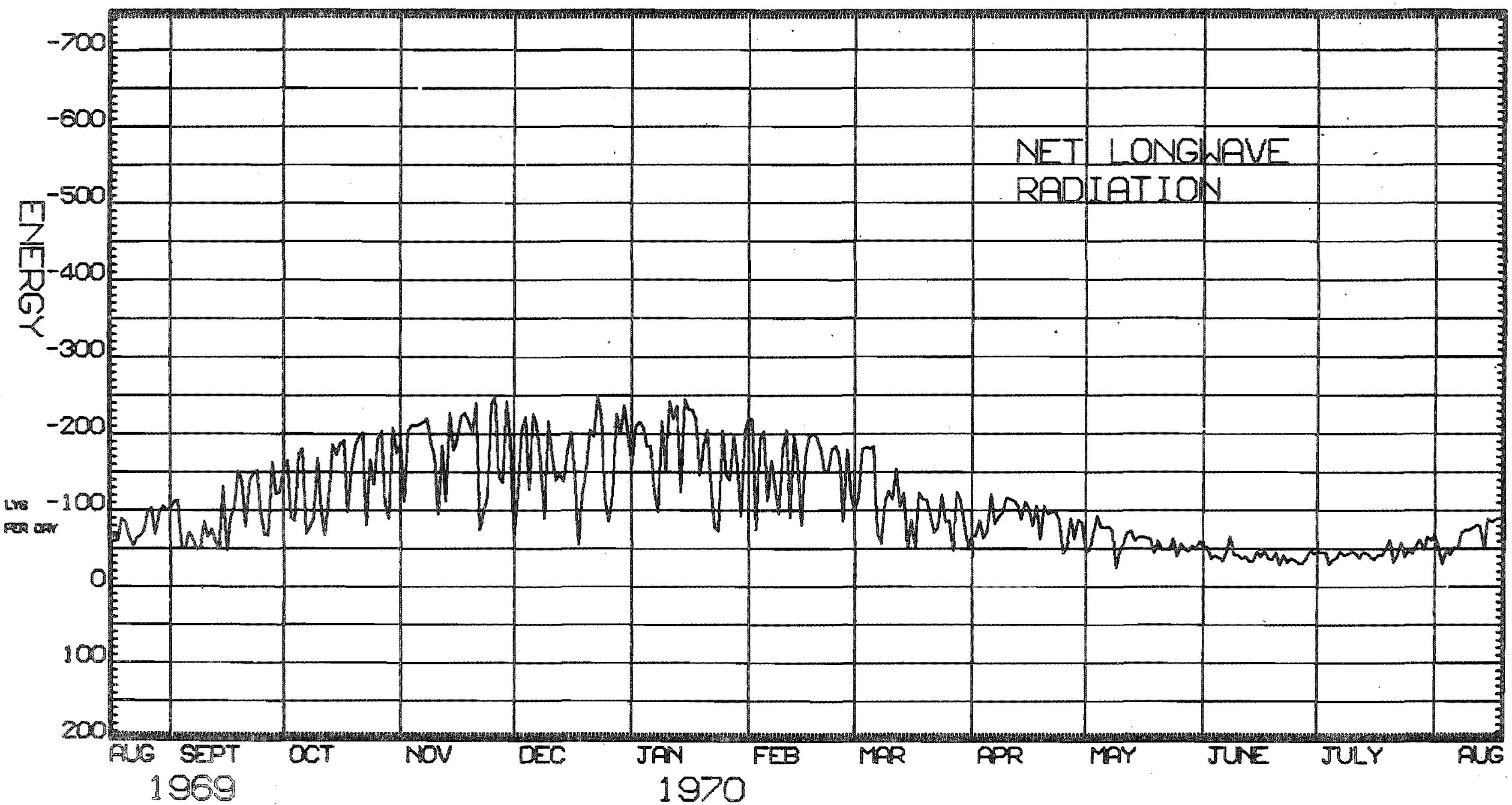


FIGURE 2.19: Daily values of net longwave radiation at the Chilton Valley during the study period

the result of other radiant flows and other factors. The second reason is to stress the need for further studies of longwave radiation at this location. The radiation balance components are examined both for individual days, and for the study period as a whole.

A resolution of the radiation balance for selected days in different seasons is shown in Table 2.10. Diffuse radiation has been measured using the method described in section 2.2. Albedo values are assumed to be those shown in Table 2.10, following the discussion and results presented in section 2.3. As an additional computation the albedo for July 22 is taken to be 0.50 representing that for snow cover. This is done for the sake of illustration since, although data on the resolution of q and Q were available for this day, no snow cover was suggested when the criteria described in section 2.4 were applied. No resolution between upward and downward longwave radiation is made due to lack of data.

The dominance in the magnitude of Q is seen on all dates. The increase of the q/Q ratio for the winter date (see section 2.2) is also apparent. Comparison between the sets of values for July 22 shows that with snow cover, the changes in the longwave, shortwave and total radiation balance values would be well marked. In the calculations used here, the relatively higher α value operates on both Rn and net shortwave radiation, and subsequently on net longwave radiation. In reality, the higher α value affects the net shortwave radiation and the net longwave radiation and then the value of Rn.

The relative importance of the net longwave radiation

TABLE 2.10

VALUES OF RADIATION BALANCE COMPONENTS FOR SELECTED DAYS IN DIFFERENT SEASONS. UNITS ARE LY DAY^{-1} EXCEPT FOR ALBEDO WHICH IS DIMENSIONLESS

Date	Q	q	α	Net LW	Net SW	Rn
Oct 20	521	49	0.13	-195	495	300
Dec 22	669	66	0.13	-247	640	393
Apr 17	257	26	0.13	-114	246	132
July 22	93	27	0.13	-59	104	45
July 22*	93	27	0.50	-44	60	16

* With assumed snow cover

term can also be seen. In the radiation balances of October 10, December 22 and April 17 net longwave radiation is, respectively 0.39, 0.39 and 0.46 of the net shortwave radiation value. In the winter example, the proportion is 0.57 or 0.74, in the case of snow cover. A similar pattern can be seen on a seasonal scale (Fig. 2.18). Here the net longwave radiant flow is about one third of the size of the net shortwave flow in summer, but in winter the two are almost comparable in size.

The above discussion helps to illustrate the composite nature of Rn. It also emphasises the importance of net longwave radiation. The relative difficulty of measuring the longwave fluxes has, in the past, often led to an underestimation of its importance. It is clear that a more complete resolution, including both upward and downward flows of short and longwave radiation would be beneficial.

2.9 Shortwave Radiation on Slopes

Since solar radiation is a significant energy source in a large variety of physical processes it is of interest to examine the shortwave radiant energy input on the different slopes that comprise the Chilton Valley. One method by which this can be approached is to compute the theoretical amounts of direct SW↓ arriving on the slopes by means of a method proposed by Garnier and Ohmura (1968). With respect to the position of the recording site, the valley may be simplified into seven major slope facets. The characteristics of these slopes are shown in Table 2.11 and they are located on Figs. 2.20, 2.21, and 2.22. The amount of direct SW↓

TABLE 2.11

MAJOR SLOPE CHARACTERISTICS IN THE CHILTON VALLEY

Slope	Descriptive Term Used in Text	Slope Angle Degrees	Slope Orientation Degrees from N
A	South east	37	150
B	Recorder site	10	225
C	NW facing	30	315
D		30	88
E	East facing	14	90
F		34	280
G	SW facing	34	220

incident on these slopes was computed for the solstices and the equinoxes. An atmospheric transmissivity of 0.80 was used. No adjustment for sky line obstruction was made. The significance of this omission is considered below. As an aid to the following discussion solar path diagrams for slopes A and C are presented in Figs. 2.23 and 2.24 in addition to that for the recorder site, slope B, already shown in Fig. 2.1.

The December direct SW↓ totals, Fig. 2.20, show that all slopes receive high amounts of radiation. The slopes on the eastern side of the valley have the greatest north-facing component and thus have potentially higher radiation inputs than those on the western side of the valley. The greatest difference occurs between the southeast facing slope (A) and the east facing slope (E) and is 231 ly day^{-1}, representing 30% of the direct SW↓ on the east facing slope. It should also be noticed that the recorder site slope (B) receives the second highest amount of direct SW↓ at this solstice.

In the periods of equinox, Fig. 2.21, the differences of direct SW↓ due to slopes are better defined. The north west facing slope (C) is now the slope with the highest direct SW↓ while the south east facing slope (A) still receives the least. The difference, of 403 ly day^{-1}, between these two now represents 74% of the direct SW↓ input on the sunnier slope.

At the time of the winter equinox the differences between the slopes are most well defined. The south east facing slope (A), according to the computational method used, now

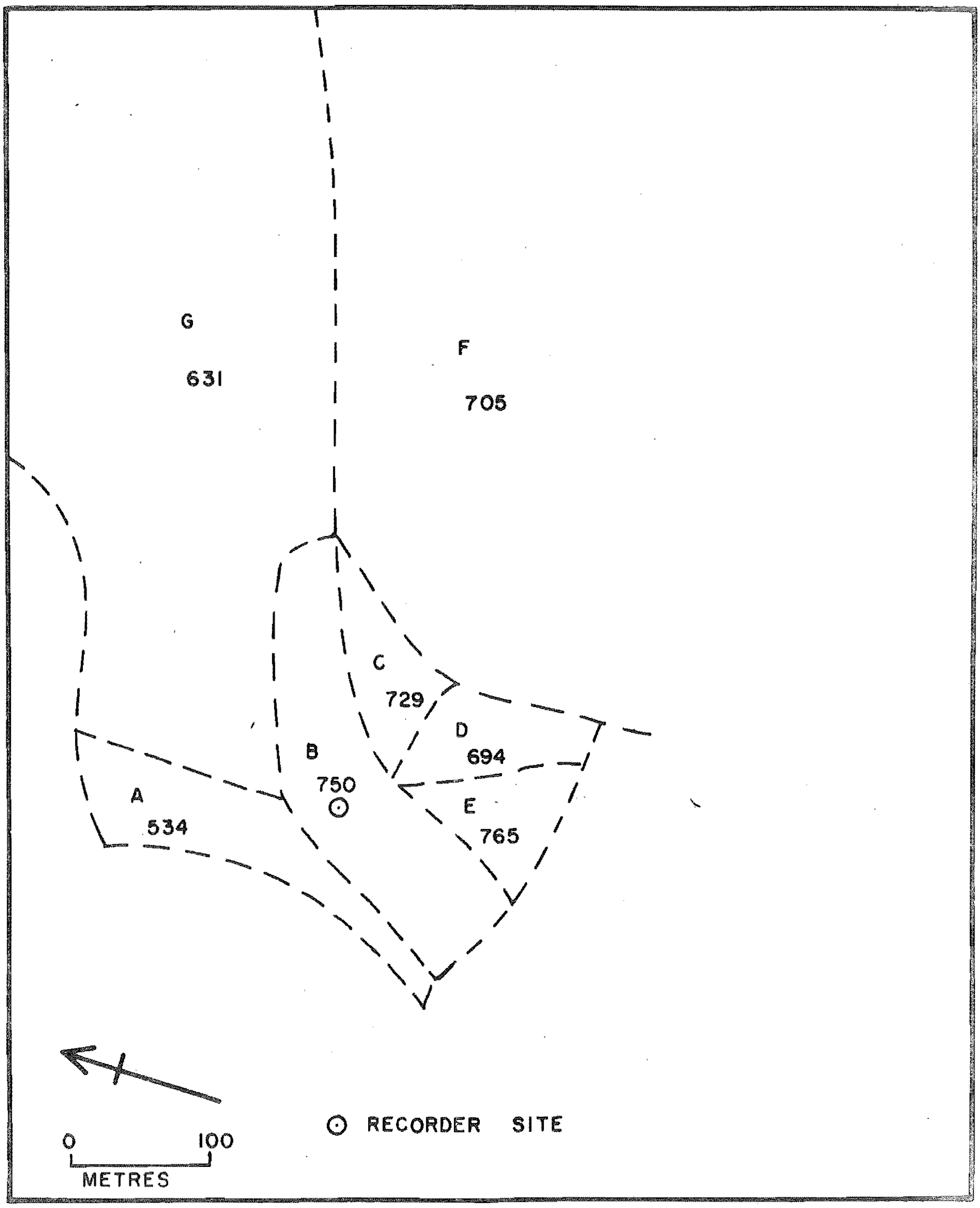


FIGURE 2.20: Potential total of direct shortwave radiation, for an assumed atmospheric transmissivity of 0.80, on slopes in the Chilton Valley on December 22 (in ly day^{-1})

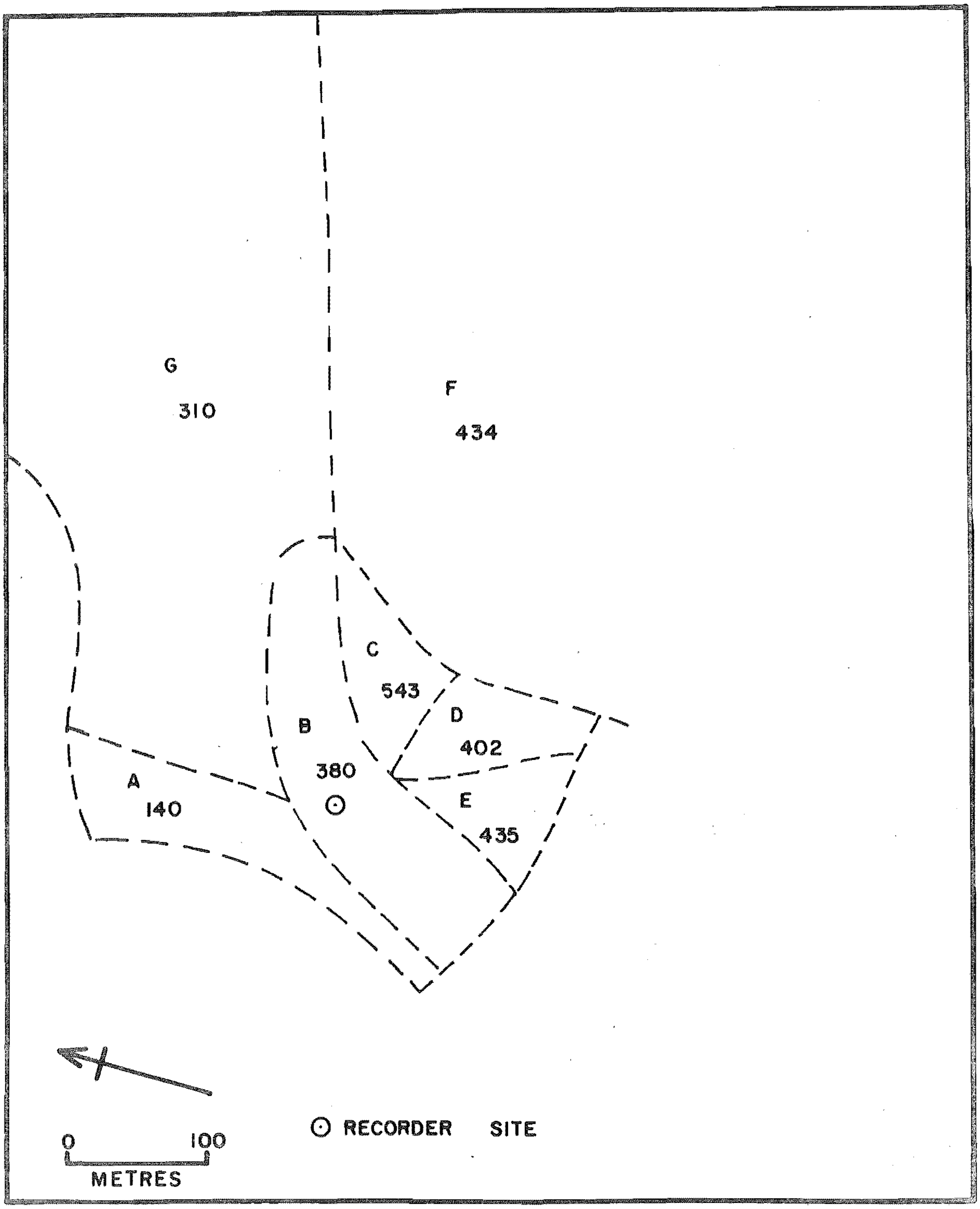


FIGURE 2.21: Potential total of direct shortwave radiation, for an assumed atmospheric transmissivity of 0.80, on slopes in the Chilton Valley on March 21 and September 23 (in ly day^{-1})

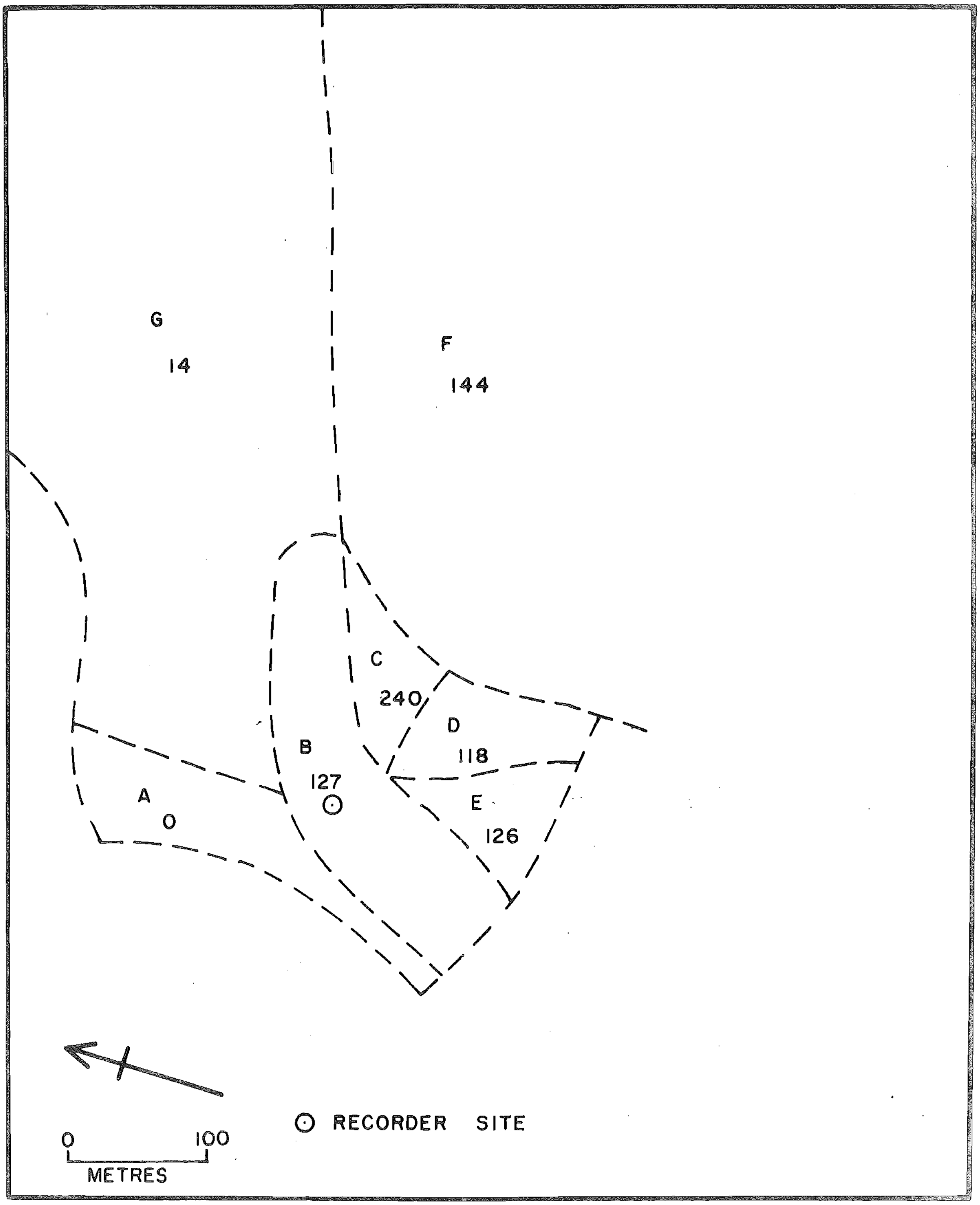


FIGURE 2.22: Potential total of direct shortwave radiation, for an assumed atmospheric transmissivity of 0.80, on slopes in the Chilton Valley on June 22 (in ly day^{-1})

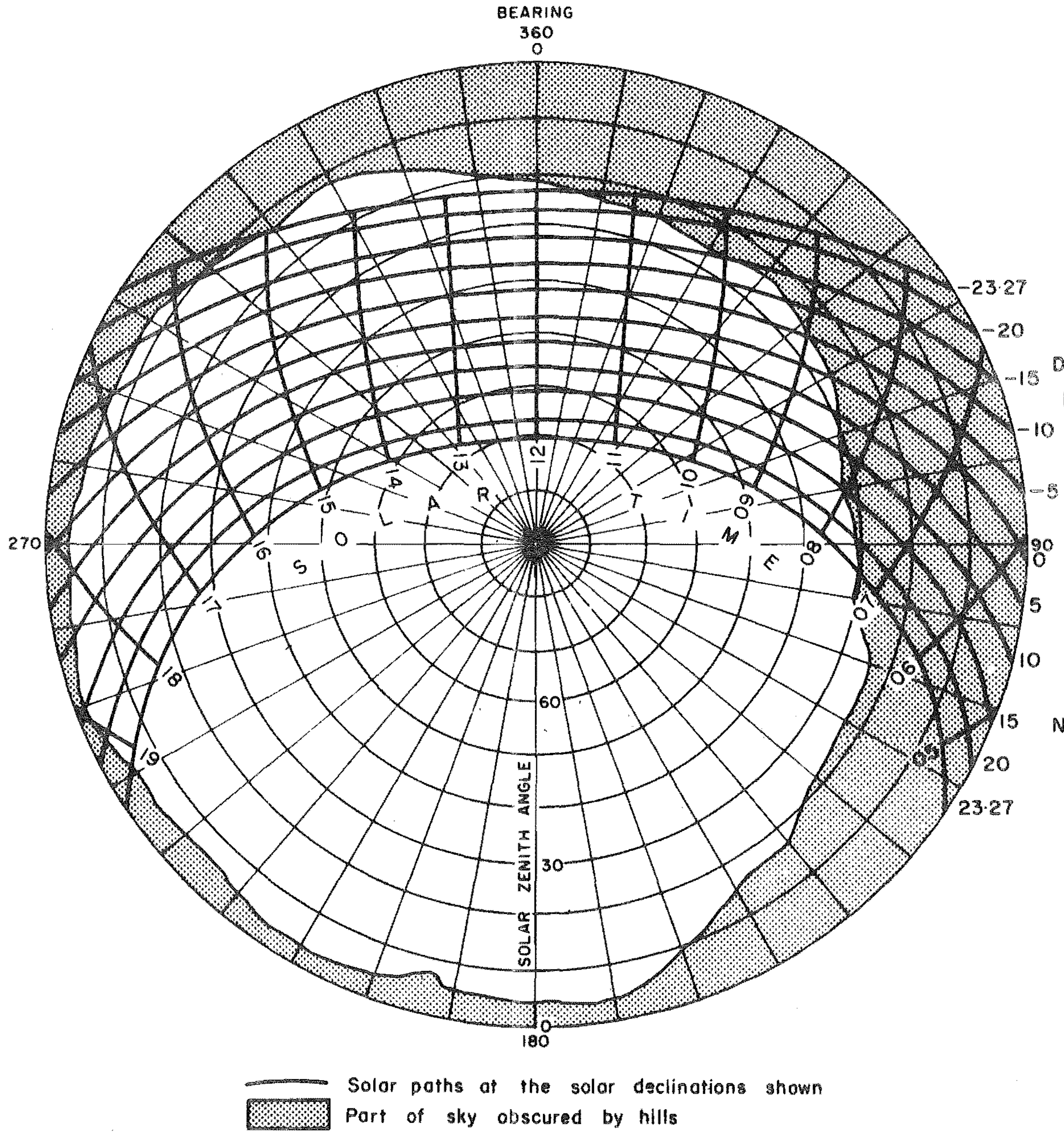


FIGURE 2.23: Solar path and horizon diagram for slope C

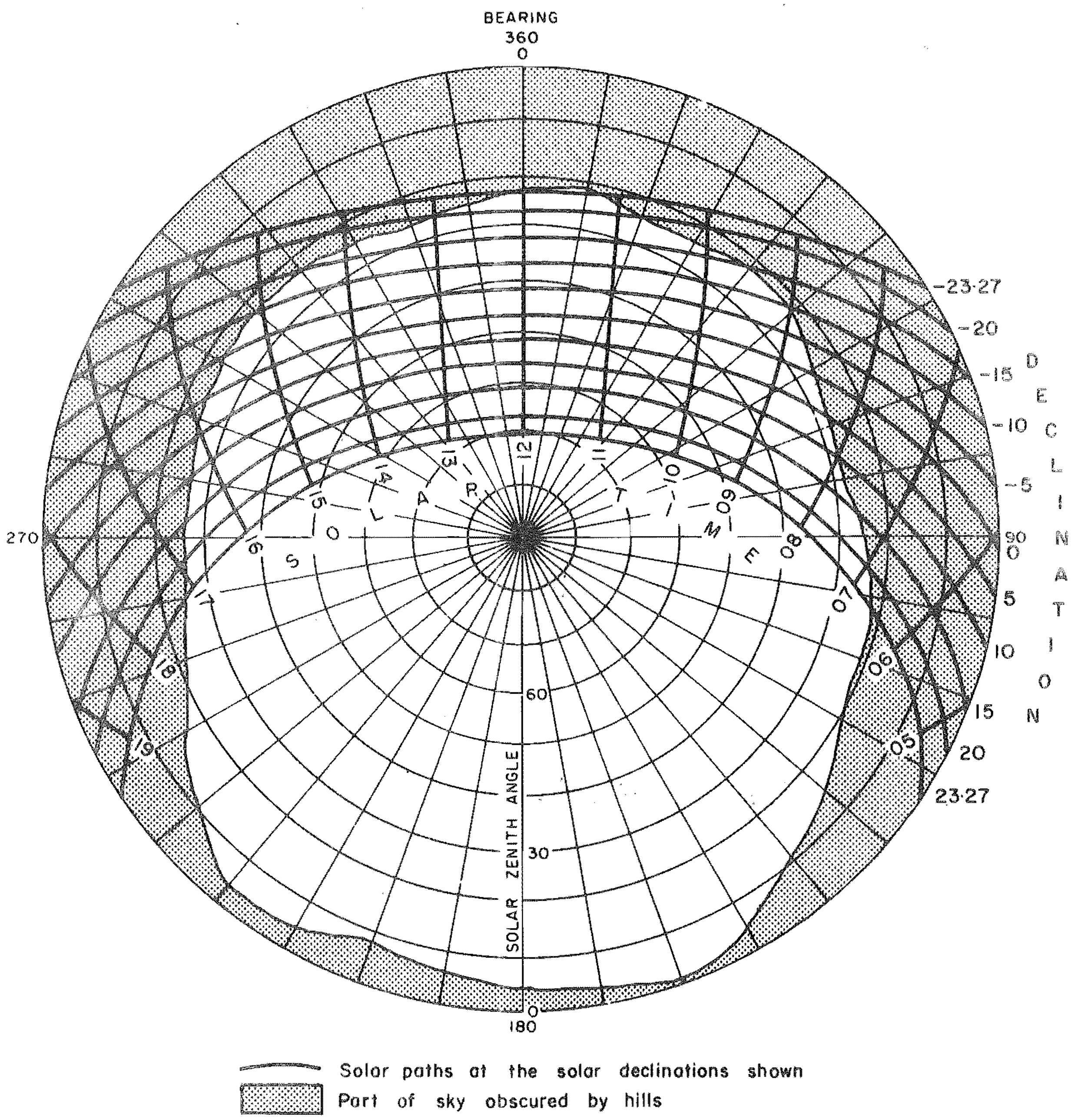


FIGURE 2.24: Solar path and horizon diagram for slope A

receives no direct SW↓ at all and the south west facing slope (G) receives only 14 ly day^{-1}. The north west facing slope (C) is again potentially the greatest receiver of direct shortwave energy.

In general, the effect of sky line obstruction in the Chilton Valley site on the above results will be greatest in winter. As was shown in Table 2.4 for a horizontal plane on the recorder site slope, during the summer months the loss of direct SW↓ was about 5%, but in the winter months this figure could be as high as 60%. Consideration of the solar path diagram (Fig. 2.23) for the north west facing slope (C) which generally had the higher direct SW↓ input, shows that this slope would undergo a significant loss of direct SW↓ only in the morning. However, the equivalent diagram (Fig. 2.24) for the south east facing slope shows that this shaded slope would lose direct SW↓ for quite long periods in both the morning and the afternoon. Sky line obstructions will act to reduce all of the direct SW↓ totals shown above, but, due to the topography of the valley, it can be seen that a potentially greater reduction will occur on the south east facing slope (A) than on the north west facing slope (C). This will therefore accentuate the differences in direct SW↓ already shown to exist as a result of slope orientation and angle. The importance of the differences of direct SW↓ on slopes suggests that further studies should be undertaken to include systematically the effect of sky line obstructions. Methods of including this factor have been suggested by Ohmura (1970) and Hay (1971).

It should also be pointed out that only direct SW↓

has been computed in this study. Jackson's (1967) comments on the importance of diffuse radiation at all seasons were mentioned in section 2.2. In addition, Samukashvili (1969b) has demonstrated that the effect of cloud cover is to equalise SW↓ on different slopes, and furthermore, that snow cover, which, in the Chilton Valley, more often occurs on the higher parts of slope F, can increase diffuse SW↓ on slopes with large inclinations.

Owing, therefore, to diffuse SW↓ , the actual differences in total SW↓ on the various slopes will not be quite as great as those reported for potential direct SW↓ in Figs. 2.20 - 2.22. However, the differences in the Chilton Valley will still be large and, as shown for other locations (Rouse and Wilson, 1969), can have a marked effect on factors such as snow melt and soil moisture. As an example of the importance of the shortwave radiant inputs on the Chilton Valley slopes, the work of Owens (1967) may be quoted. He found that mass movement was mainly associated with freeze-thaw action, and that most mass movement occurred on the north west facing slope (C). If it is assumed that in order to give rise to a thaw it is necessary to receive direct SW↓ for a period longer than ten minutes, then comparing the north west facing slope (C) with the south east facing slope (A), only the former could experience thawing at the time of the winter solstice. General observation confirms that the freeze-thaw process is greatest on the north west facing slope (C) and the rapid disappearance of ice from this slope during the day is frequently noted (Soons and Rayner 1968).

2.10 Summary

The most important points arising from this study of the radiation heat flow in the Chilton Valley may be summarised as follows:-

1. High day to day variabilities in shortwave, net shortwave, net longwave and net radiation are observed.
2. During the study year the highest monthly SW↓ values occurred in November and January, and not in December as might have been expected. Comparison with other data indicates that the high SW↓ value in November was unusual. The possibility of such a variation in monthly SW↓ is important, as it could have a marked effect on other energy flows at the location.
3. There is evidence that the Chilton Valley has less SW↓ in spring and early summer, than is recorded at Christchurch, although for the rest of the year recorded values are similar. Since, on the basis of a decrease of atmospheric mass with altitude, the Chilton Valley might be expected to receive more SW↓ than Christchurch, it is probable that a relatively greater average atmospheric moisture content occurs over the valley, especially in the spring and early summer. Precipitation data are in accord with this suggestion.
4. Measurements of albedo for surface cover in the valley ranged from 10.2% to 18.9%, with the value for tussock cover being about 13%. During the winter with snow cover, or partial snow cover, albedo values proved to exceed 50%.

5. Examination of the relation between shortwave radiation and net radiation daily totals shows that, with the data available at present, the most satisfactory equation for deriving daily net radiation totals is

$$Rn = 0.65\ (1 - \alpha)\ SW\downarrow - 22.8\ \text{ly day}^{-1} \qquad 2.5.12.$$

The large increase in the value of albedo with snow cover is the reason for the inclusion of the albedo term which otherwise may have been omitted.

6. Mean monthly net radiation during the study year ranged from 269 ly day^{-1}, in November, to 9 ly day^{-1}, in June. Although there was a positive radiation balance for the recorder site in winter in the period studied, the year to year variation in SW$\downarrow$ implies that this need not always occur.

7. Studies of the direct SW$\downarrow$ reception on slopes indicate that some slopes in the valley would have a negative mean monthly radiation balance in winter.

8. The composite nature of net radiation has been emphasised, and attention has been drawn to the relatively large size of net longwave radiation and the need for further study of its components.

9. Differences in direct SW$\downarrow$ received on different slopes in the valley are most well marked in winter. There is evidence that this has an effect on freeze-thaw cycles, and subsequently on mass movement processes.

C H A P T E R T H R E E

SOIL HEAT FLOW AND RELATED PHENOMENA

3.1 Introduction

The quantity of heat flowing into and out of the soil is normally much smaller in value than that of the other major components of the surface heat balance equation (equation 1.1.1.). However, within the wider context of the physical processes of the earth's surface, the movement of heat within the top layers of the soil is important for three reasons. Firstly, these layers are fundamental for vegetation growth. Secondly, it is here that most geomorphological processes perform their work. Thirdly, A is important because it is involved in a certain amount of feedback into the surface energy exchange system. Since both vegetation growth and, to a slightly lesser extent, geomorphological processes are affected by the amount of heat present at any one time, the relatively small magnitude of the soil heat flow term in the heat balance context may give a false impression of its importance. As an example of the feedback mechanism, the amount of soil heat flow can affect the quantity of $LW\uparrow$. Further, together with soil moisture conditions, A can influence the quantity and partitioning of heat loss by means of the evaporative and sensible heat flows.

In order to illustrate the relation of soil heat flow to other phenomena occurring in the soil, attention is not only given to the amounts of soil heat flow, but also to its relation to the wider topics of the effect of moisture addition to the soil and the growth of needle ice. In this chapter, some of the thermal properties of the soil are discussed before dealing with the quantities of soil heat flow and soil temperatures. A study of the effect of rainfall on soil heat flow illustrates some of the relations between these two parameters. Finally, attention is turned to a geomorphological agent that is especially important in the valley - needle ice. The microclimatic conditions pertaining to needle ice growth are examined within the context of the surface energy exchange.

In order to avoid possible confusion, it is appropriate to state here that in the following sections flow of heat <u>away</u> from the surface, and therefore into the soil, is regarded as <u>negative</u>. This convention is being adopted in order that compatibility with the rest of the present study which deals with the surface of the earth, is maintained. If soil heat flow were being considered by itself, it would be more fitting to use the opposite signs.

3.2 Thermal Properties of the Soil

Some of the physical characteristics of the soil have been described in section 1.4. The thermal properties of soils in general are notoriously variable and depend on soil moisture content, mineral and organic content, temperature (where latent heat terms enter), compaction, and depth. However, in order that some comparison can be made with other locations it is thought worthwhile to attempt to calculate some of the thermal properties of the soil at the recorder site in the Chilton Valley. The values of thermal conductivity relate to certain days which are specified and used in studies of the effect of rainfall on the soil (section 3.5).

Values of thermal conductivity (k) were estimated using the standard formula

$$A = - k\ dt/dz \qquad 3.2.1.$$

where A, the heat flow through the layer, was approximated by the records of the heat flux plates, and the temperature gradient dt/dz, was taken between the thermistors at 0.5 cm and 2.0 cm depths. The calculations were made for data between 0200 hr. and 0600 hr., as this is when nearly steady cooling at night permits a pseudo steady state formula to be used (Brooks,1959). In fact values of k were calculated for all periods during the 24 hours, and the above time showed the greatest stability in k values.

The k values obtained (Table 3.1) compare favourably with examples quoted by Brooks (1959) and Van Wijk (1965), but some, especially that of 1 June 1966, are outside the range of 1.91 - 3.87 m cal cm^{-1} $^{o}C^{-1}$ found at Taita, New Zealand by Gradwell (1968). The value of k for water is relatively high, and this is illustrated by the value for the rainy day of 5 October, and the slightly higher value for 15 February. Although the possibility of instrumental error cannot be discounted the latter value demonstrates the sensitivity of k to additional water, since, up to the time when k was measured, only 0.08 cm of rain had fallen. Also of note is the high value of 1 June when sub-zero temperatures occurred. This high value may be attributed to the fact that the thermal conductivity of ice is about four times that of water. Fig. 3.10 indicates that during the period for which k is computed, the 0^oC isotherm reached down to 2.0 cm, therefore implying lower temperatures above this level. Allowing for some depression of the freezing point of the soil moisture, it is probable that some of the soil moisture in the 0.0 - 2.0 cm level would have been in a solid state at this time. The change to a liquid state on June 2 gives rise to a value of k comparable with that of the wet October 5.

The heat capacity (C) of the soil is more regular and more predictable than the thermal conductivity. Using data on the proportion of mineral and organic material in the soil, supplied by the University of Canterbury Botany Department (C.J. Burrows, pers. comm.), a sample calculation of the volumetric heat capacity was computed by means of the formula

$$C = 0.46x_m + 0.60x_o + x_w \qquad \text{\_\_\_\_ 3.2.2.}$$

In this formula, which is given by de Vries (1963), x_m, x_o and x_w are the volume fractions of the mineral matter, organic matter, and water respectively. At the Chilton Valley, this gave values of C ranging from 0.47 cal cm^{-3} $^oC^{-1}$ for a completely dry soil or one that contained ice (when the coefficient of x_w is 0.46) to 0.63 cal cm^{-3} $^oC^{-1}$ when the soil contained 30% water. These values may be compared with 0.3 cal cm^{-3} $^oC^{-1}$ for dry sand or clay, and 0.7 cal cm^{-3} $^oC^{-1}$, for sand or clay with a 40% water content (Van Wijk,1965). The Chilton Valley computations indicate that although there is some variability in heat capacity with changes of moisture

TABLE 3.1

VALUES OF THERMAL CONDUCTIVITY, k, AND TEMPERATURE (AVERAGED BETWEEN 0.5 AND 2.0 cm LEVEL) BETWEEN 0200 AND 0600 HRS FOR SELECTED DAYS. UNITS OF k ARE $m.cal.\ cm^{-1}\ sec^{-1}\ {}^{o}C^{-1}$

Date	4 Oct 1966	5 Oct 1966	13 Feb 1967	15 Feb 1967	1 June 1966	2 June 1966
k	3.33	5.65	3.25	3.32	13.30	5.68
temp °C	4.1	6.4	12.6	11.6	-0.1	3.3

TABLE 3.2

SOIL HEAT FLOW IN LANGLEYS (AVERAGE OF TWO SENSORS, NEGATIVE AWAY FROM SOIL SURFACE

	Total for the Month	Daily Average
August	306	18
September	655	22
October	-40	-1
November	-402	-13
December	-278	-9
January	-372	-12
February	74	3
March	96	3
April	207	7
May	104	3
June	55	2
July	117	4
August	110	7

and/or ice content, the total range is not very large.

3.3 The Amount of Soil Heat Flow

In August 1964 three soil heat flux plates were inserted at a depth of 1.0 cm in close proximity to the net radiometer site. One placed under a tussock (*Festuca novae-zelandiae*) showed virtually no heat flow, and was soon re-sited so that there were two, F_1 and F_3, under *Raoulia subsericea* and *Agrostis tenuis*, and one, F_2, under *Raoulia subsericea* and small stones. Thus the soil heat flow measurements are typical not of the relatively large tussock plant, whose large mass prohibits soil heat flow measurable with the present methods, but of some of the smaller plants that live in association with the tussock. The second flux plate, F_2, was removed in July 1967 for laboratory ice needle experiments. The results below represent the average of the two remaining plates. It is probable that this sample, of two plates, is not statistically large enough, but further equipment and information were unavailable to expand the study. Soil heat flow investigations are frequently performed using three or less transducers (Davies *et.al.*, 1969; Terjung *et.al.*, 1969; Thompson *et.al.*, 1971). In the present case it is impossible, without more data, to assess the extent to which the measured values are representative of the surface of the site. Further attention should be directed towards the sampling problem in this location where heterogeneity of surface cover is the norm.

The accuracy of the measured values is also enigmatic. The accuracy of the plates as indicated by the manufacturer's calibration is 5%. Calculations from soil temperature data (section 3.4) suggest that the measured daily totals are low, although the results from these calculations are not directly comparable for reasons given in the next section. The problem of the accuracy of the soil heat flow values is also discussed in appendix B. Overall, it is thought that the soil heat flux plates give useful records of the relative day to day changes of soil heat flow, but that absolute monthly totals derived from them may be underestimates. Fortunately, the size of the soil heat flow term is usually much less than that of the other heat balance components, and on a long term basis it is often neglected (e.g. Budyko, 1958).

Variation of soil heat flow through the study period (Table 3.2 and Fig. 3.1) reveals an interesting pattern. Large quantities of heat left the soil between August and early October. This possibly results from the relatively large inputs of heat following high amounts of radiation in the previous summer and autumn, and low surface and air temperatures during the above months. There follows a relatively short transition period, in October, between outflow and inflow of soil heat. Subsequently, in November, the month of highest solar radiation, the maximum value of heat input into the soil is recorded. The lack of radiant input in December is clearly reflected in the soil heat flow values. Then, following relatively high inflows in January, there begins a long transition period, between inflow and outflow, covering the period between late January to early March (see Fig. 3.2). Autumn and winter monthly totals are positive, but not large,

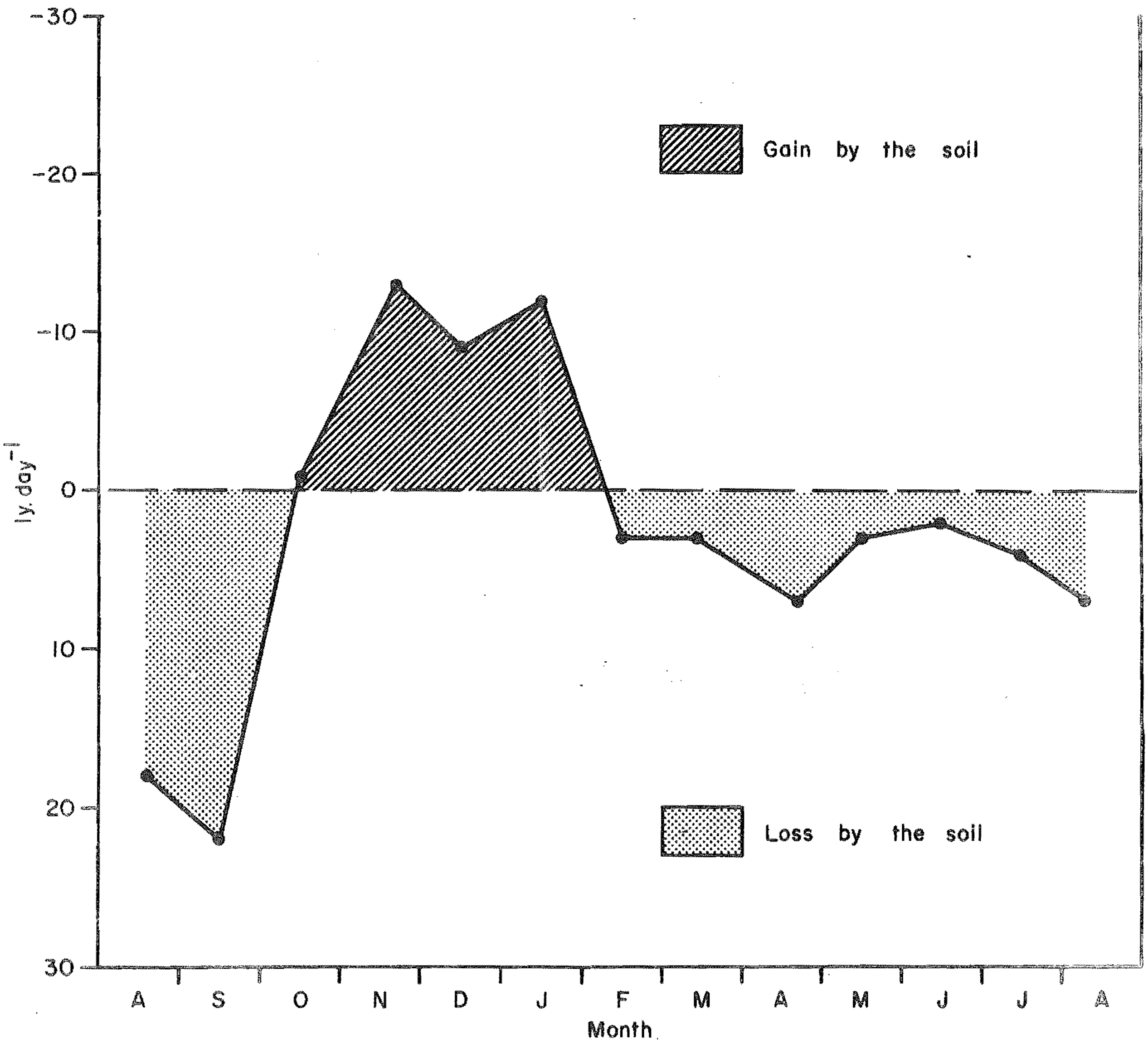


FIGURE 3.1: Monthly mean values of soil heat flow for the Chilton Valley during the study period in ly day^{-1}. Heat flow away from the surface is negative.

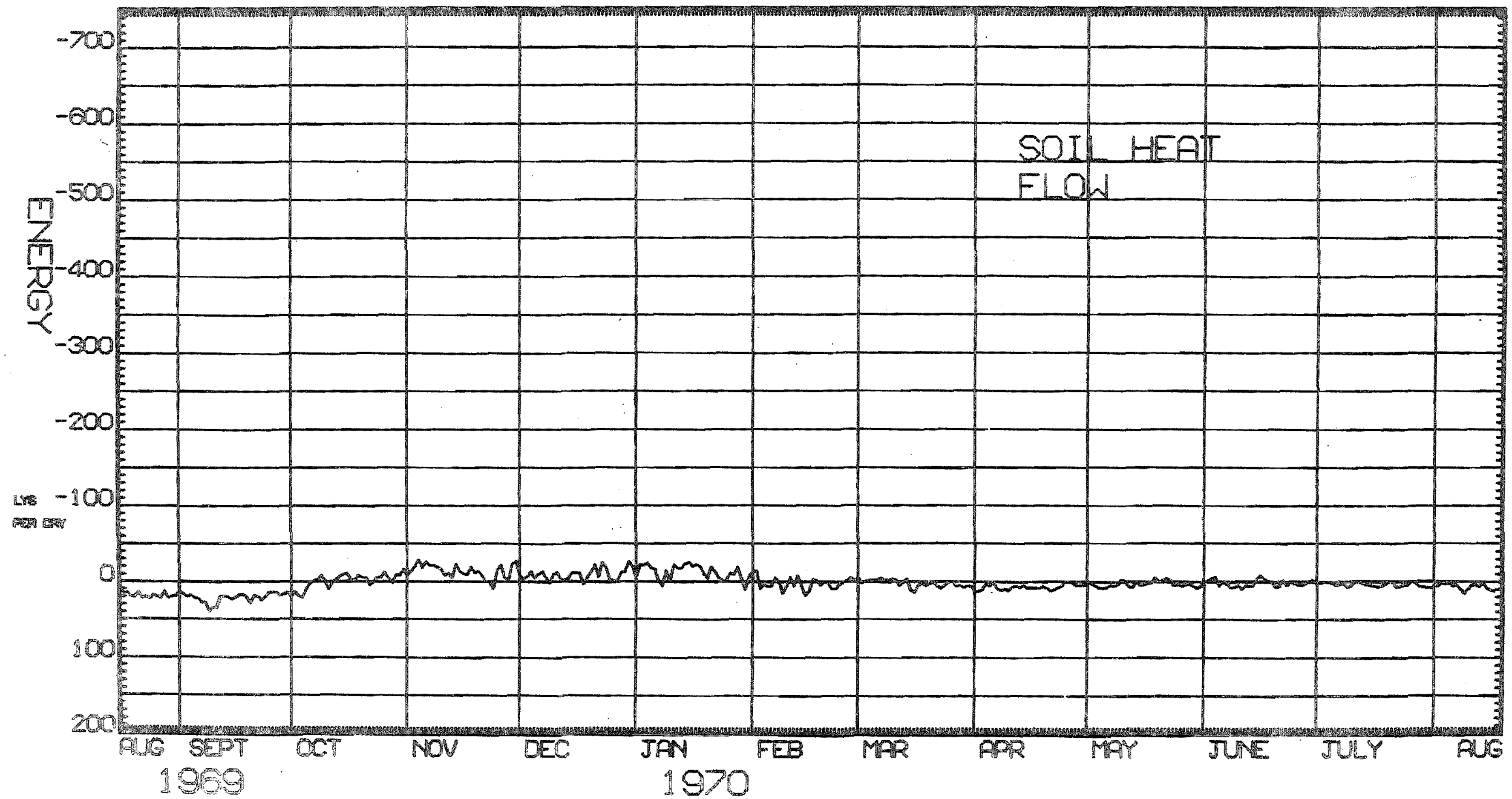


FIGURE 3.2: Daily values of soil heat flow for the Chilton Valley during the study period. Heat flow away from the surface is negative.

and a fairly close balance is reached between the heat input in the summer months, 1092 ly, and the output, 863 ly, in the winter months that have been recorded. Possibly, if records for succeeding months were analysed, the remaining balance might be found. Monthly A totals recorded here are in the same order as those found by Frankenberger (1962) in Schleswig-Holstein, where a variation between -400 ly month^{-1}, in January to 430 ly month^{-1}, in May, was reported. The data in Table 3.2 indicate that during the summer months, when there is high radiant input and daily net soil heat flow is directed away from the surface, the value of this latter flow is sensitive to changes in the radiative flux. This sensitivity is not so well marked at times of low radiation values, such as winter, or at times when the direction of net monthly A is changing its sign.

The same is true when the daily data (Fig. 3.2) are examined. In the summer months there is little or no lag between days of high Rn and high A values directed away from the surface. The linear regression between A and Rn daily totals in November, December and January is

$$A = 0.021\ Rn - 26.8 \qquad \text{3.3.1.}$$

C.C. = 0.73 S.E.E. = ± 3.4 ly day^{-1}

However, when the data for the whole year are combined the equation is

$$A = 0.019\ Rn - 29.0 \qquad \text{3.3.2.}$$

C.C. = 0.57 S.E.E. = ± 5.4 ly day^{-1}

Again it appears that at times of relatively 'steady state' flow of heat into the soil, there is a fairly good relationship between A and Rn.

The remainder of the variance in the relationship would probably be explained by factors such as soil moisture, which will affect P and LE, and, consequently, the amount of heat available for passage into the soil. Nevertheless, the relationships above, in part demonstrate the feedback mechanism mentioned in section 3.1.

The daily values of A (Fig. 3.2) show a large amount of day to day variability. In summer it is quite possible on overcast or rain-days to have an outflow of A, while conversely, in winter, it is possible to record an inflow. The early months of the study period are an exception to this, but even here there are relatively large variations in the actual positive values of A. This day to day variability is not surprising, bearing in mind the relation of A to Rn, and the variability of the latter, as already shown in section 2.6. Since soil moisture conditions can also affect A (see section 3.5) the degree of daily soil heat flow variability can be readily understood.

3.4 Soil Temperature

Soil temperatures are closely related with soil heat flow, but the relationship is not always simple owing to time and space variations in the thermal conductivity and diffusivity of the soil. Soil temperatures were measured at three depths, and in the manner described in section 1.5. However, some difficulty was experienced in obtaining continuous records at the 15.0 cm and 30.0 cm level, since junctions between the thermistors and the connecting cables proved susceptible to penetration by water. This difficulty was not experienced during the summer months. Data for the 2.0 cm level were available for all of the year, but data for the 15.0 cm and 30.0 cm level were available only between 7 November to 13 May, and 7 November to 25 March respectively. In the analysis, air temperature records for every hour were used but for soil temperatures, only daily maxima and minima were extracted from the original chart recordings.

Monthly mean data (Fig. 3.3 and Table 3.3), where available, show that higher temperatures generally occur at the higher soil levels, but in April temperatures at 15.0 cm have a higher average value than those at 2.0 cm. Maximum temperatures at all levels, including screen level, occur in January, lagging some two months after the peaks in Rn and soil heat flow input. However, a lag of two months is not necessarily normal since the double maxima of Rn in November and January, and the relatively low value in December, are probably atypical (see sections 2.2 and 2.6). The available data suggest that the annual temperature cycle is seen at all levels examined. The average monthly temperatures at all levels appear to be compatible with the expected changes, i.e. higher temperatures at higher levels in summer, and lower temperatures at higher levels in winter (Unger, 1951). Although winter temperatures at the lower levels are not shown here, a changeover point appears to occur in May, when the mean temperatures at the 2.0 cm and 15.0 cm levels are the same. In April, as mentioned, the higher temperature occurs at the 15.0 cm rather than the 2.0 cm level. That the Chilton Valley temperatures should follow the expected pattern is interesting in that soil temperatures recorded at the

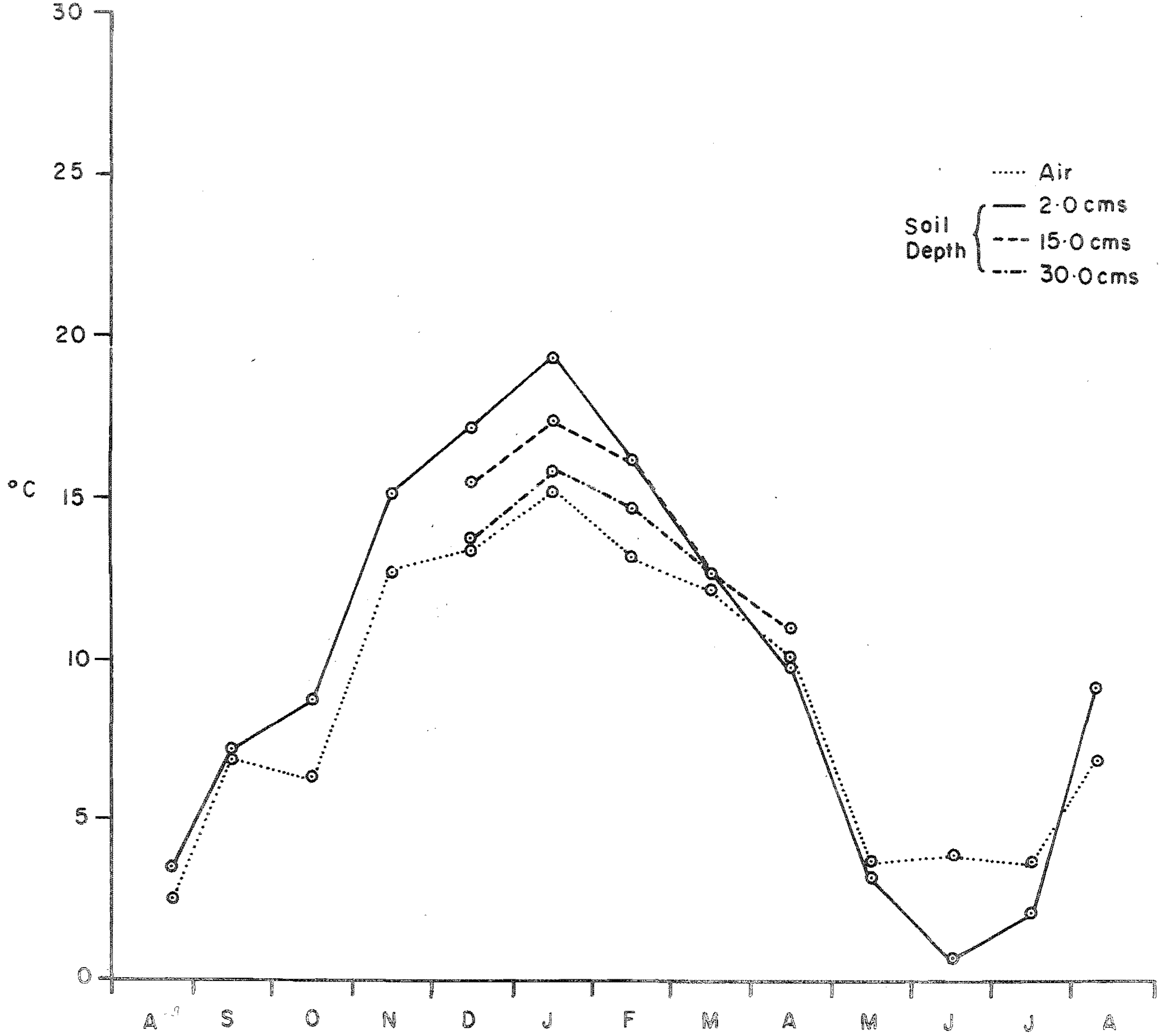


FIGURE 3.3: Monthly mean values of soil and air temperature at the Chilton Valley during the study period

TABLE 3.3

MEAN MONTHLY SOIL AND AIR TEMPERATURE IN DEGREES CENTIGRADE

VALUES ARE NOT SHOWN WHERE DATA ARE INCOMPLETE OR MISSING

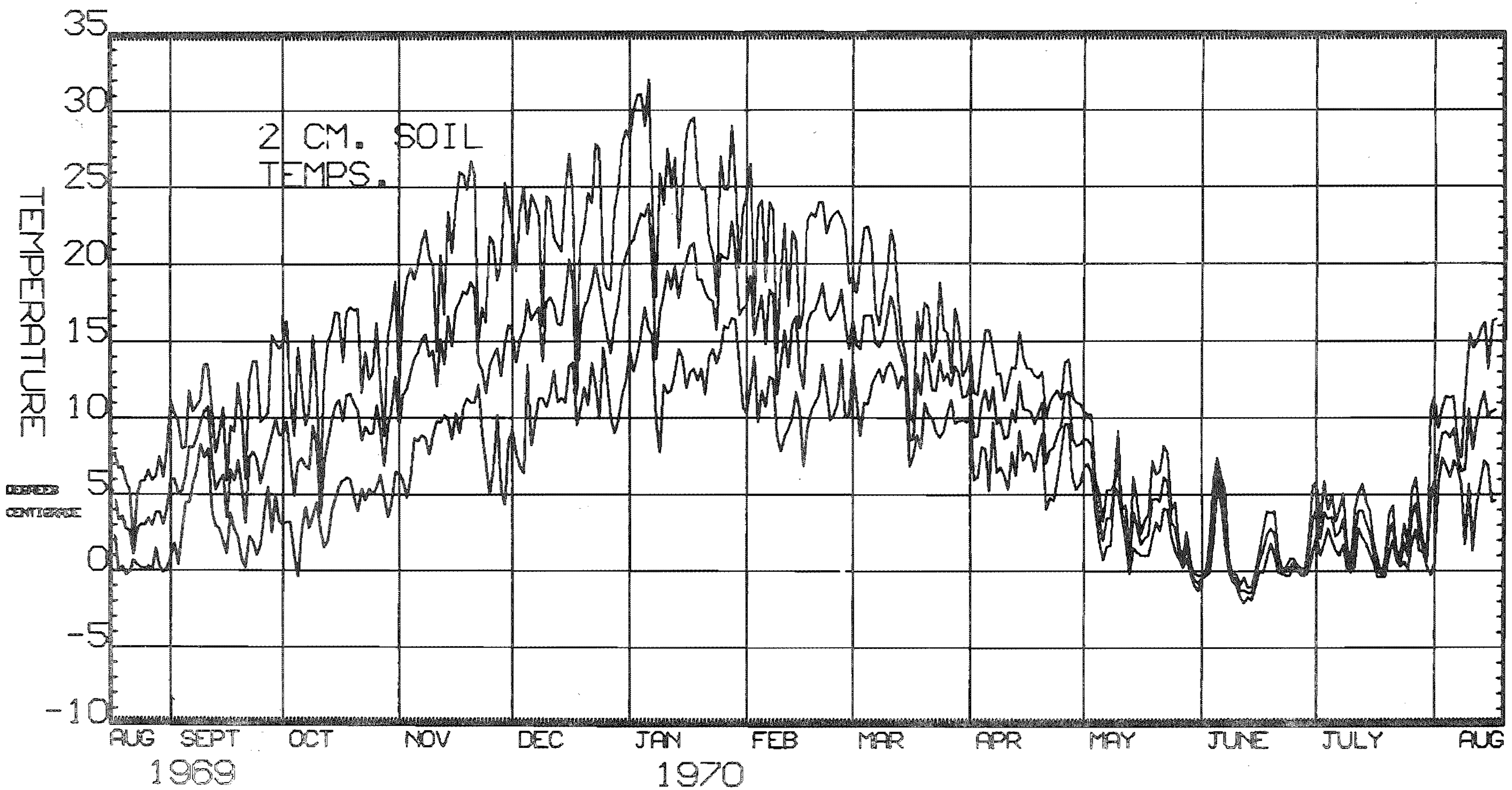
		Aug.	Sept.	Oct.	Nov.	Dec.	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Year
Air	Mean	2.7	6.9	6.3	12.8	13.4	15.1	13.1	12.1	10.0	3.7	3.9	3.7	6.9	8.8
	Max.	8.1	10.8	11.5	20.5	18.4	20.6	19.7	16.7	15.0	8.2	6.6	6.5	10.3	13.6
	Min.	-0.9	3.4	2.2	5.7	9.2	10.3	7.3	8.4	5.9	0.1	1.1	1.1	2.0	4.6
2 cm	Mean	3.5	7.1	8.8	15.1	17.2	19.4	16.1	13.6	9.8	3.1	1.5	2.2	9.1	10.0
	Max.	6.4	10.9	13.5	21.0	23.1	25.5	21.4	16.6	12.6	4.3	2.2	3.7	12.7	13.7
	Min.	1.5	5.5	4.5	4.4	7.1	8.1	5.3	6.4	4.0	0.1	0.4	1.0	5.5	4.2
15 cm	Mean					15.5	17.4	16.0	13.6	10.9					
	Max.					16.8	18.7	17.2	14.4	11.5					
	Min.					14.3	16.1	14.9	12.9	10.3					
30 cm	Mean					13.6	15.8	14.6	12.6						
	Max.					13.8	16.0	14.9	12.7						
	Min.					13.4	15.6	14.4	12.5						

nearby Craigieburn Station, at 990 m show, in some years, lower monthly mean temperatures at higher levels for almost every month of the year (N.Z.M.S.,1970).

The daily values of mean, maximum and minimum temperatures at the 2.0 cm level are shown in Fig. 3.4. Once more a high degree of day to day variability emerges. Also interesting are the wide ranges (up to almost 20°C) that can occur between maximum and minimum temperatures in summer, although in the winter the ranges are quite small. In fact, on the average, the temperature range between maximum and minimum temperatures varies between 11.3°C in January and 5.4°C in July. Daily mean, maximum and minimum soil temperatures at the 15.0 cm and the 30.0 cm levels are shown in Figs. 3.5 and 3.6 respectively. Soil temperatures at 15.0 cm show smaller extremes between daily maxima and minima, but daily changes are still clearly marked at this level. At 30.0 cm the difference between maxima and minima is almost always less than 1.0°C. Furthermore, daily changes are not as obvious as at the higher levels, but changes in the order of a week stand out well.

The simultaneous variations of daily mean temperatures at the 2.0 cm, 15.0 cm and 30.0 cm levels are shown in Fig. 3.7. The effect of synoptic scale weather changes (see section 7.3) is clearly shown in data for the summer months. During anticyclonic situations, temperatures at all levels gradually rise for several days, but when a well marked southerly cold front occurs there is a sudden fall in the temperatures at 2.0 cm and 15.0 cm. A temperature decrease at 30.0 cm is also noted, but a lag of about two days occurs

FIGURE 3.4: Daily values of soil temperature at 2 cm depth at the Chilton Valley during the study year. Daily values of maximum temperature (top line), mean temperature (middle line), and minimum temperature (bottom line) are shown.



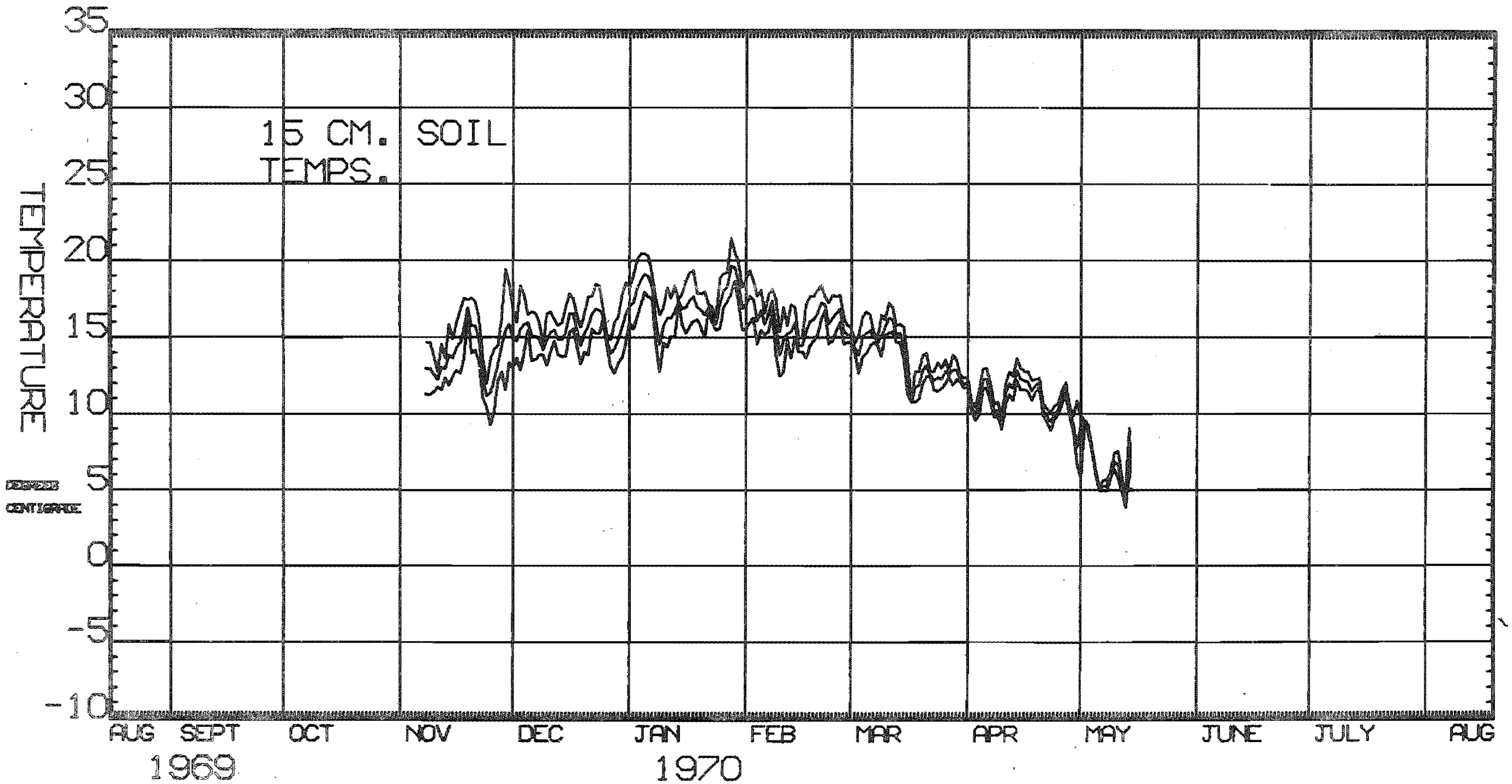


FIGURE 3.5: Daily values of soil temperature at 15 cm depth at the Chilton Valley for part of the study year. Daily values of maximum temperature (top line), mean temperature (middle line), and minimum temperature (bottom line), are shown.

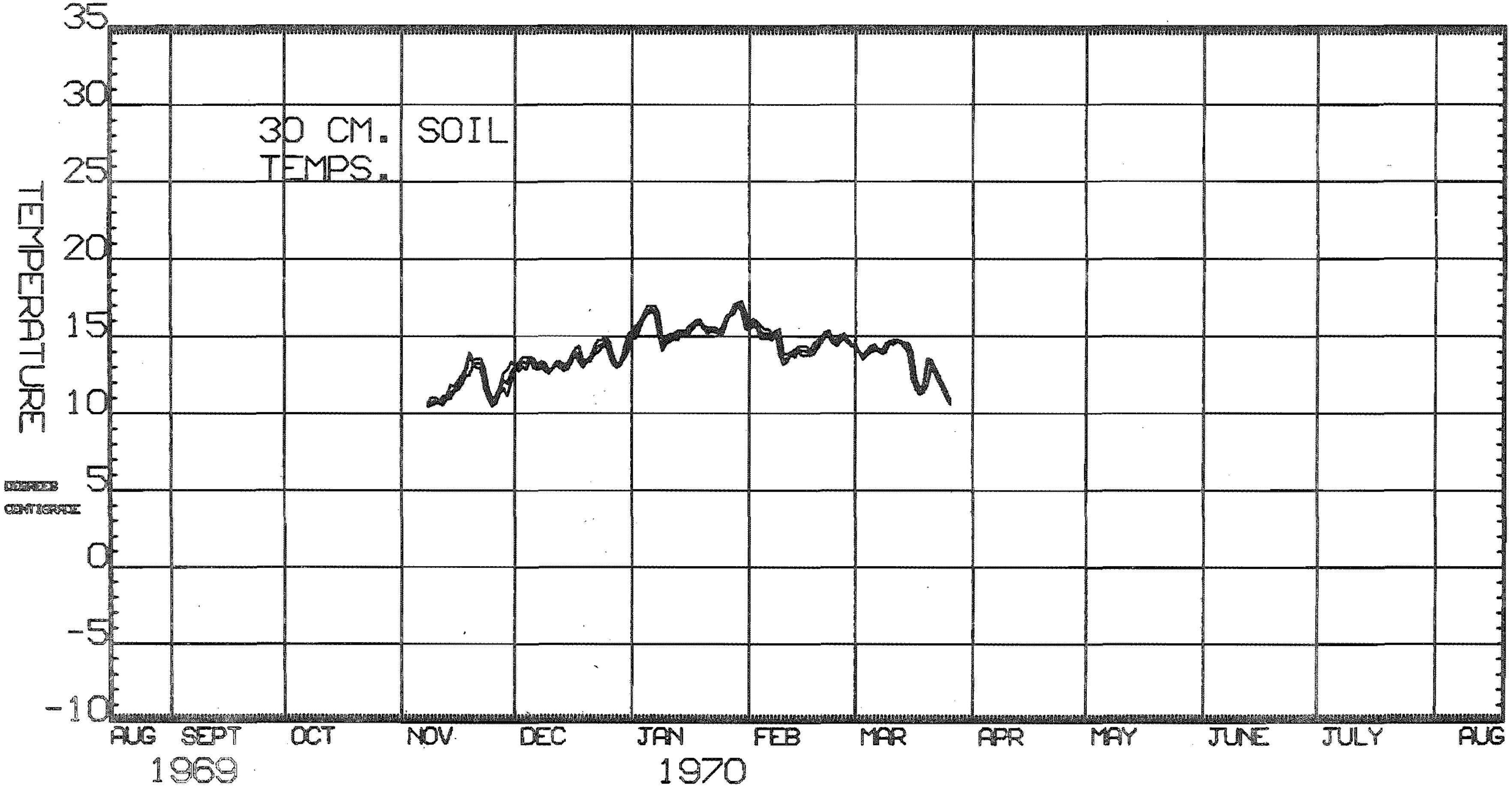


FIGURE 3.6: Daily values of soil temperature at 30 cm depth at the Chilton Valley for part of the study year. Daily values of maximum temperature (top line), mean temperature (middle line), and minimum temperature (bottom line), are shown.

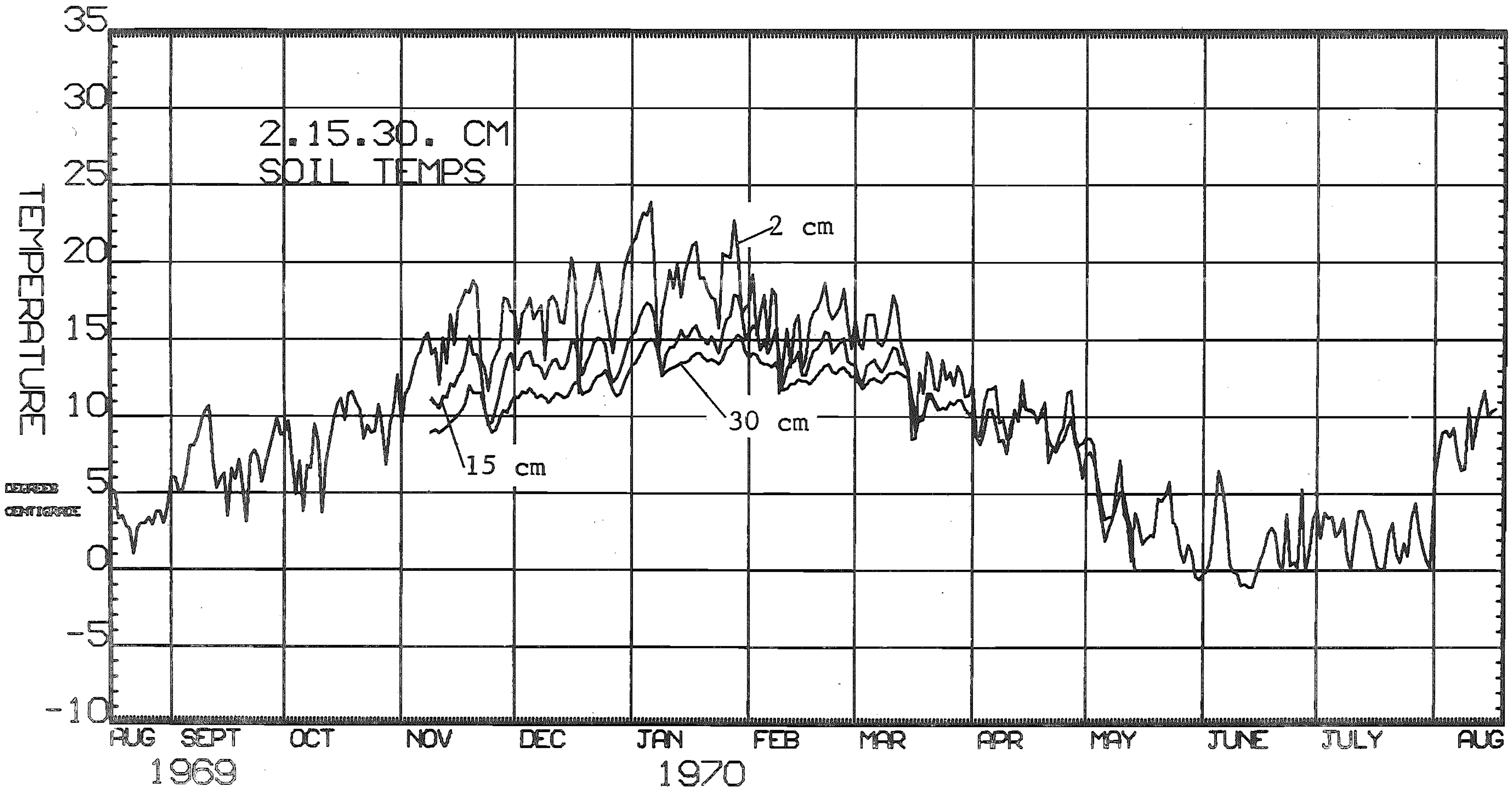


FIGURE 3.7: Mean daily values of soil temperature at 2 cm, 15 cm, and 30 cm depth at the Chilton Valley for part of the study year.

at this level. Good examples of the effect of frontal passages are seen for the data of 22 November and 7 January. Similar relationships, but referring to weather singularities, are reported by Unger (1951).

Information on the frequency of freeze-thaw cycles can be obtained from the records of air temperature and soil temperature at 2.0 cm. For the purpose of this analysis a freeze-thaw cycle is here defined as a period when the minimum temperature is less or equal to 0.0°C, and the maximum temperature for the following day is greater or equal to 0.0°C. This definition is used due to the availability of the data extracted from the original charts. The number of consecutive days below freezing is the number of successive days when both maximum and minimum temperatures are less or equal to 0.0°C. As might be expected, the winter months show the highest frequency of freeze-thaw cycles and days of continuous freezing (Table 3.4). The air thermistor shows higher frequencies of freeze-thaw cycles than the 2 cm soil thermistor, but there are no consecutive days when screen temperature is below zero. In the the study period there are only four complete months during which no days with sub zero screen temperatures occur. If the normal variation of temperature with height (see for example Lowry, 1969) is followed, it may be expected that the frequency of freeze-thaw cycles at the surface, in this location, will be even greater than that at screen height. If this is so, the present records confirm the conclusion of Morris (1965), that in this area, ground frosts can occur at any time of the year.

As mentioned at the beginning of this section, soil

TABLE 3.4

FREQUENCY OF FREEZE-THAW CYCLES AT THE RECORDER SITE DURING THE STUDY PERIOD

	2 cms		Air (Screen)	
	Freeze-thaw Cycles	Consecutive days Below 0^oC	Freeze-thaw Cycles	Consecutive days Below 0^oC
August	5	0	12	0
September	0	0	4	0
October	1	0	6	0
November	0	0	1	0
December	0	0	0	0
January	0	0	0	0
February	0	0	0	0
March	0	0	0	0
April	0	0	1	0
May	1	4	18	0
June	9	6	10	0
July	2	2	10	0
August	0	0	0	0

temperatures are related to the flow of heat through the soil. It is of interest to compare the soil heat flow indicated by the monthly mean soil temperatures with that measured by the soil heat flux plates. A thermal conductivity of 3.25 m cal cm^{-1} sec^{-1} ^{o}C is assumed as being representative of a dry day in summer (see section 3.2). Application of equation 3.2.1. for the temperature gradient between 2.0 cm and 15.0 cm gives monthly mean A values of -36, -43, -22, 0, and +24 ly day^{-1} for December through to April respectively. These values are approximately four times larger than the values given by the soil heat flux plates. However, they cannot be taken as being necessarily more representative than the flux plate recordings for the following reasons. Firstly, they refer to the 6.5 cm level whereas the flux plates refer to the 1.0 cm level; although in this situation the higher level might be expected to show higher daily totals of A. Secondly, the temperature based calculation assumes the soil to be homogeneous, and the thermal conductivity, calculated for the 0.5 cm to 2.0 cm layer, to apply to the 2.0 cm to 15.0 cm layer. Thirdly, the temperature profile is taken about 70 cm from the flux plates. Although some other workers (e.g. Thompson, et.al., 1971) have found similar difficulties in relating flux plate records to soil temperature profiles, it is concluded here that the flux plate records may possibly underestimate daily totals. The degree of underestimation must await the collection of further data.

3.5 The Effect of Precipitation on Soil Heat Flow and Soil Temperature

Diurnal variations of soil heat flow and soil temperature were examined in detail in a study of the effect on these parameters of precipitation following a period of dry weather. Three periods of two days each, in spring, summer and winter, were selected. Each period contained a day, at the end of a dry spell, with a relatively large amount of SW↓, followed by a day with rain. The days chosen, and other relevant information are shown in Table 3.5. The days precede the dates of the main study period.

In order to examine soil temperature changes, isotherms on a time-depth section, and temperature profiles for specified times, were drawn. The section for the 4-5 October, 1966 (Fig. 3.8), shows a typical temperature pattern. Both the heating and the cooling waves can be seen by looking, in particular, at the 15^{o}C and 5^{o}C isotherms. The smaller waves on the second day when it rained, are clearly demonstrated. The time lag of both heating and cooling, as greater depths are sampled, is also apparent. The temperature profiles at four-hour intervals may be used to show approximately those layers of the soil that are heating and cooling at these times, and also to estimate, by extrapolation, the actual surface temperatures. Extrapolation of the profiles of 4 October gives surface temperatures of approximately 1^{o}C and 27^{o}C for 0400 hr. and 1200 hr. respectively. The times indicated at the top of the profile diagrams are those when isothermal temperatures prevailed in the 0.5 - 2.0 cm layer, thus showing the changeover points from heat outflow to inflow,

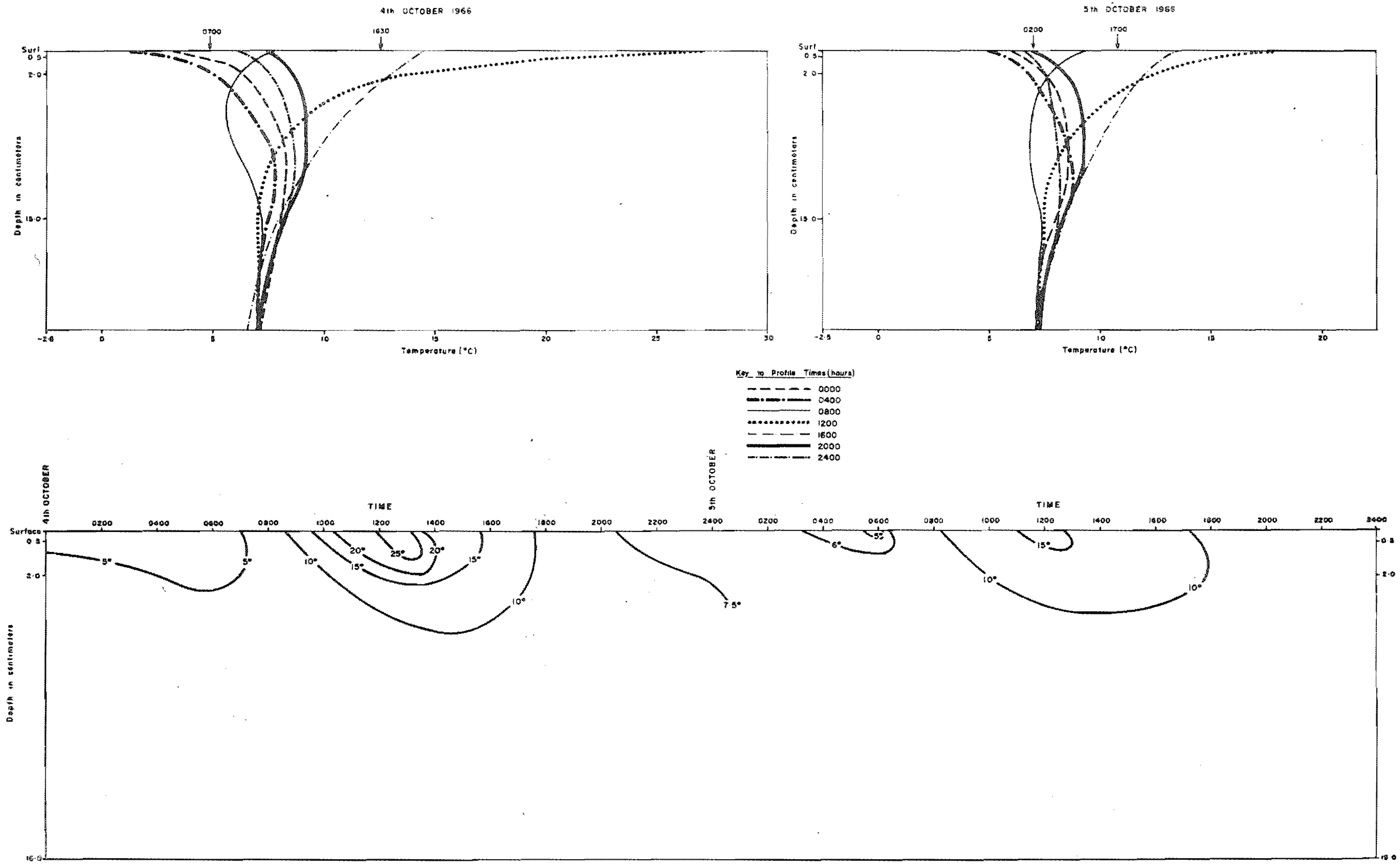


FIGURE 3.8: Soil temperature profiles and time-depth sections for 4 and 5 October 1966.

and the reverse, within this layer. The convergence of the profiles in the deeper layers illustrates, once more, the decrease of daily temperature fluctuations with depth. The difference between the profiles for the 4 and 5 October indicates that on the cooler, wet day fluctuations were less at all levels.

A decrease in temperature extremes is also manifest in the summer example (Fig. 3.9). Here however, is seen an outstanding contrast between the two days in the temperature gradients, and the size of the heating waves. The temperature gradient for the 1600 hr difference between 0.5 cm and 10.0 cm on 13 February is as large as 17.5 ^{o}C i.e. 1.84 ^{o}C cm^{-1}. Extrapolated surface temperatures for 1600 hr and 2400 hr respectively, range between approximately 37 ^{o}C and 8 ^{o}C, a change of 29 ^{o}C in eight hours. Even higher surface temperatures have been found at other times. On 30 January 1966, for example, at 1600 hr, a surface temperature of approximately 42 ^{o}C (107.8 ^{o}F) was extrapolated. The time section for 13 February also shows the deep penetration of the summer heat. In comparison, the wet day, 15 February, shows a much smaller penetration. The profiles and sections for this day are noteworthy for the relatively long period of pseudo-isothermal conditions that are shown between about 0700 hr and 1300 hr. It should be mentioned that despite the rain the profiles, in particular, are comparable to those of the fine spring day (4 October), and in fact Rn on these two days was very similar. Although this could be due to coincidence, this implies that, for these examples, soil temperature conditions may be influenced by rainfall but are

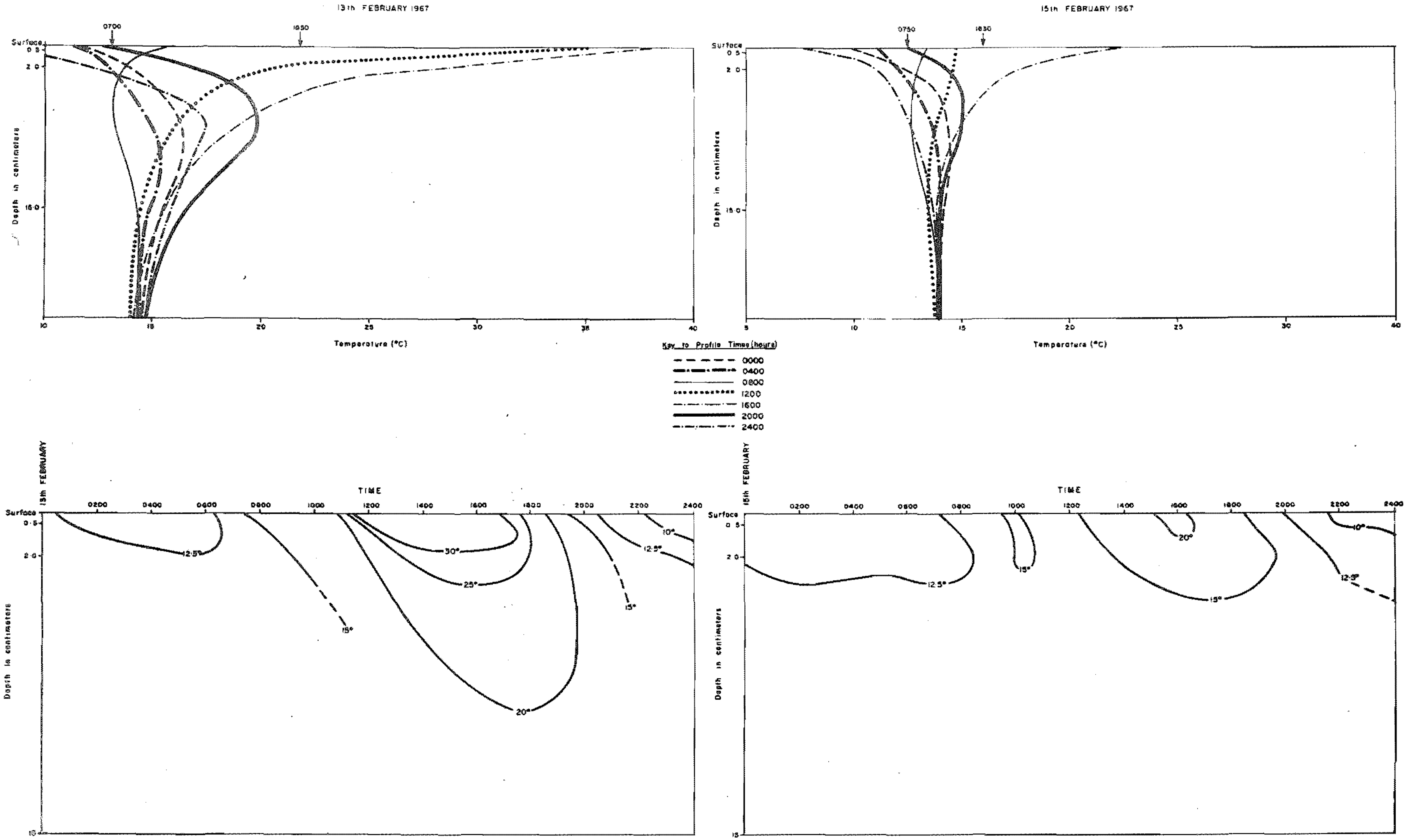


FIGURE 3.9: Soil temperature profiles and time-depth sections for 13 and 15 February 1967

more closely related to radiant input at the surface.

The winter examples (Fig. 3.10) provide a complete contrast to those above. In the winter case, after a relatively long, cold, dry spell, the top layers of the soil are frozen, and temperature profiles are markedly isothermal. On 1 June, when no rain fell, the temperature at 0.5 cm remained below 0.0°C until noon. The maximum temperature reached at this level during the whole day was only 2.5°C at 1500 hr. The profiles and the sections for this day indicate very little spatial or temporal temperature variation. Instead of a decrease in heat penetration, as in the previous examples, the following rainy day brought almost continuous heating to most parts of the soil. On 2 June at 15.0 cm, the temperature rose throughout the day from 1.2°C, at 0000 hr, to 4.0°C, 24 hours later. Although surface cooling occurred from about 1300 hr, the gradients on the temperature profiles, which are no longer isothermal, indicate heat flowing into the soil until 2030 hr. The section shows that the 5°C isotherm reached a depth of approximately 11 cm. This heating is probably due to the influence of the rainwater penetrating the soil and possibly, in part, to the downward flow of sensible heat from an incoming warm, moist, air mass from the northwest.

The influence of the rain water was probably the most significant factor in giving rise to the warming of the soil. The occurrence of advection of warm air is indicated by the combination of the facts that air temperature did not decrease from 1700 hr on 1 June, until 1100 hr on 2 June, and that the average air temperatures on these two days rose from 2.8°C

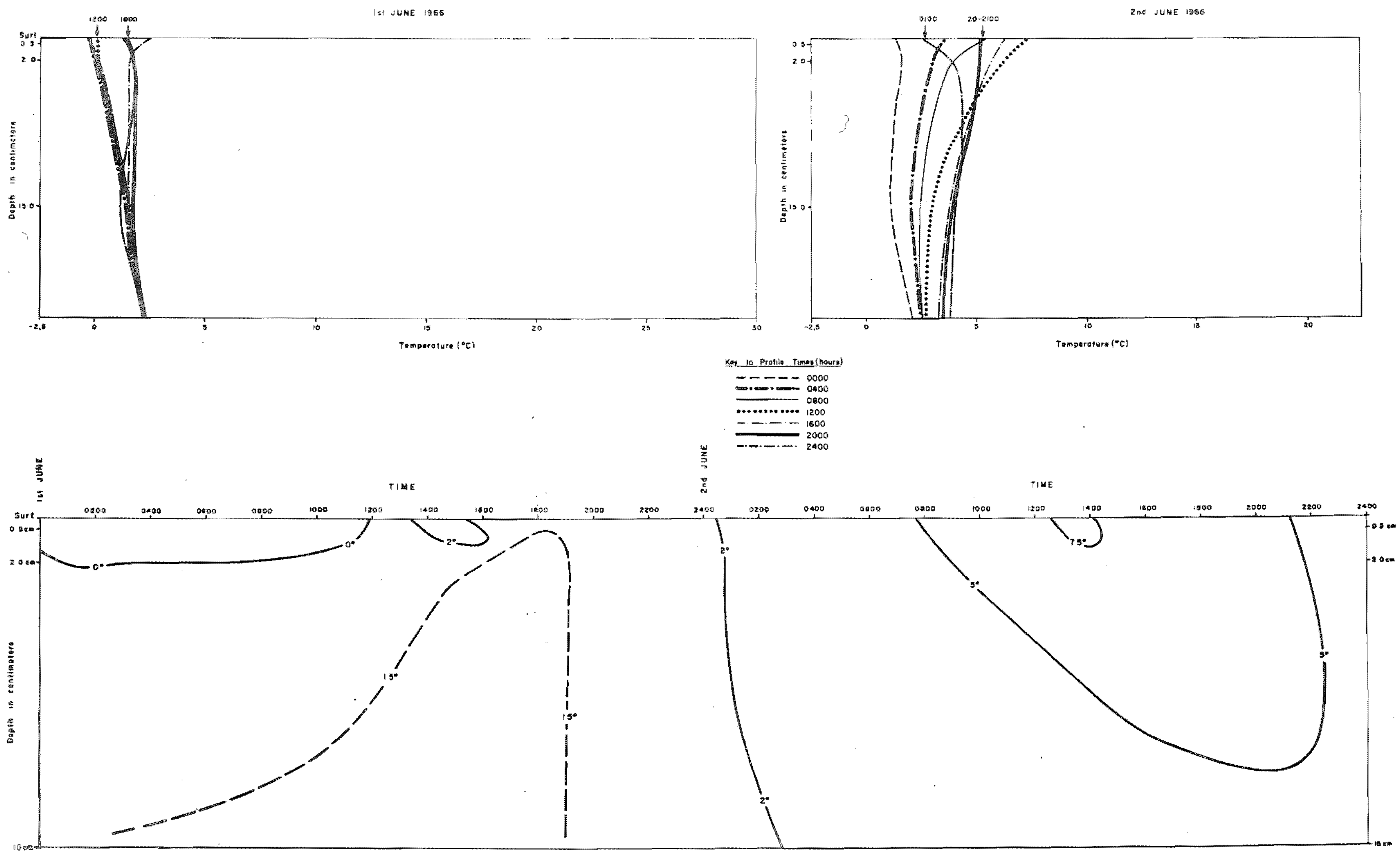


FIGURE 3.10: Soil temperature profiles and time-depth sections for 1 and 2 June 1966

to 4.5°C, despite a decrease in Rn from +40 ly day^{-1} to +21 ly day^{-1}. However, the average air temperatures, at 12 m and 3 m, for 2nd of June were 4.40°C and 4.36°C respectively, indicating that the temperature gradient, directing sensible heat downwards from the advected warm air, was not great.

Two features arise out of the above discussion. Firstly, there is a well marked reaction in soil temperatures to synoptic scale weather changes that are accompanied by rainfall. Secondly, the nature of the change can be completely reversed between periods in the winter, when the ground is already frozen, to periods in other seasons, when rainfall leads to cooling. The difference between the winter example and the summer and spring examples is also seen in the changes of thermal conductivity of the soil (section 3.2) and has been examined previously (Greenland, 1969) by means of harmonic analysis. In the context of heat balance studies the difference between the summer and spring cases and the winter case may be expressed as follows. In the summer and spring examples, although rainfall has some effect, the soil temperatures are primarily controlled by radiative input, while in the winter example, when the radiative input is already small, other factors, such as rainfall, can have an overriding effect.

Besides the examination of soil heat flow and soil temperatures, the surface heat balance for the days studied was also calculated (Table 3.6). This was done using assumptions quoted elsewhere (Greenland, 1969) which lead to an overestimation of LE especially on the dry days, with a

TABLE 3.5

PERIODS USED DURING THE PRECIPITATION-SOIL HEAT FLOW STUDY

Period	No. of dry Days Before Period	Time of Start of Rain	Total Rain During Sample Time
4, 5 Oct. 1966	10	0200 hrs 5th	1.80 cm.
* 13, 15 Feb. 1967	10	1600 hrs 14th	2.51 cm.
1, 2 June 1966	7	1000 hrs 2nd	1.60 cm.

* 14 February was omitted as this was a cloudy but almost rainless day.

TABLE 3.6

VALUES OF THE HEAT BALANCE COMPONENTS OF THE SURFACE IN cal cm^{-2} day^{-1}

Date	Rn	LE	P	A
4 October	168	-83	-82	-3
5 October	85	-83	- 8	+6
13 February	242	-182	- 8	-52
15 February	169	-160	-16	+7
1 June	40	-30	-11	+1
2 June	21	-40	+29	-10

reciprocal error in P, which was taken as the residual in the heat balance equation.

The values for 4 October show the heat balance for a dry spring day. Rn was partitioned almost equally between LE and P, and a relatively small amount of heat went into the soil. On 5 October there was a net soil heat flow toward to soil surface probably because the rain caused cooling in the top layers, and heat moved up from the lower levels. P is still negative but the value is probably small compared with the overestimated evaporative heat flow. The pattern for 13 and 15 February is similar, although more extreme. P is again underestimated, but perhaps the most outstanding aspect here is the large quantity of heat that passed into the soil on the 13 February. This amounts to over one fifth of Rn. The great thermal penetration shown in Fig. 3.9 is well accounted for by this large A value, which, it is interesting to note, was not exceeded in the long term study period. The energy balances of the surface for 1 and 2 June are in striking contrast to those of the spring and summer cases. Here the dry day, characterised by higher level soil temperatures below $0^{o}C$ for much of the time, and a small positive A value, is followed by a day of relatively large heat flow into the soil, and apparently downward transfer of sensible heat from the air. The size of this apparent positive input is probably too large, and again results from a probable overestimation of LE. Also, with the low Rn value and low soil temperature values that prevailed on these days, the input of heat due to precipitation was probably not insignificant. Therefore, the reaction of the flow of heat through the soil is similar to that of soil temperatures, and is shown to differ between the winter example and the spring and summer cases of precipitation following a dry period. In the winter example, in particular, the contrast between the two days is eloquent of the changing macro scale weather conditions.

3.6 Energy Exchange and the Growth of Needle Ice

It has been shown above that many of the thermal conditions of the soil in the winter are quite distinctive. There is considerable evidence (Gradwell, 1957; 1960, Soons, 1967; Owens, 1967) that freeze-thaw action, in association with the growth of needle ice, is a dominant agent in the erosion and enlargement of unvegetated surfaces in the South Island High Country. In other areas needle ice has also been found to be damaging to plant materials (Brink et.al., 1967). Therefore it is important to examine the surface energy exchange during the occurrence of this phenomenon. Detailed investigations were made both in the field and in the laboratory and more comprehensive information on instrumentation and procedures, than is reported here, has been published previously (Soons and Greenland, 1970).

On July 8 1967 two sites were selected for observation. These were approximately 3 m apart and beside the stream which runs close to the main recording site. Records of ground air temperatures at the main site were taken at 7½ minute intervals. Air temperatures and humidities at 2 m and at 2 cm above the ice needles were recorded at half hourly intervals (Figs. 3.11 and 3.12). A gradual decrease in temperature with an accompanying increase in relative humidity is seen. While ice needles grew at both sites, at the first site the needles grew to between 15 cm and 18 cm in length and were two tiered. At the second site the needles grew to 4 cm and were single tiered. The difference in length was probably due to the availability of soil moisture at the two sites, but the reason for the different numbers of tiers is unknown. In both sites the growth rate was observed to be higher at the beginning and decreased with time. Similar measurements of ice needle growth and meteorological factors were made on June 28, 1967.

In the laboratory, an 85 cm deep column of soil from the Chilton Valley, was reconstructed in an insulated wooden box 1 m deep and 25 cm square. The column was placed in a controlled growth chamber in the University of Canterbury Botany Department. Because of the insulation heat could only pass through the top surface of the column. Air temperature was kept at $-4.0^{o}C$ and the soil was saturated at the beginning of the experiment. Five cycles of needle ice growth were achieved without adding further water. At the end of each cycle the needles were melted by heat from an infra red lamp. During the experiment evaporation was measured by weighing two dishes, one containing water and one containing saturated soil. The results from these lysimeters are believed to be satisfactory during the freezing periods, although a slight underestimation is possible due to problems such as the dishes not having a reservoir of warm soil beneath them, and owing to the frozen soil moisture having a larger relative surface area. Evaporation rates for the heating

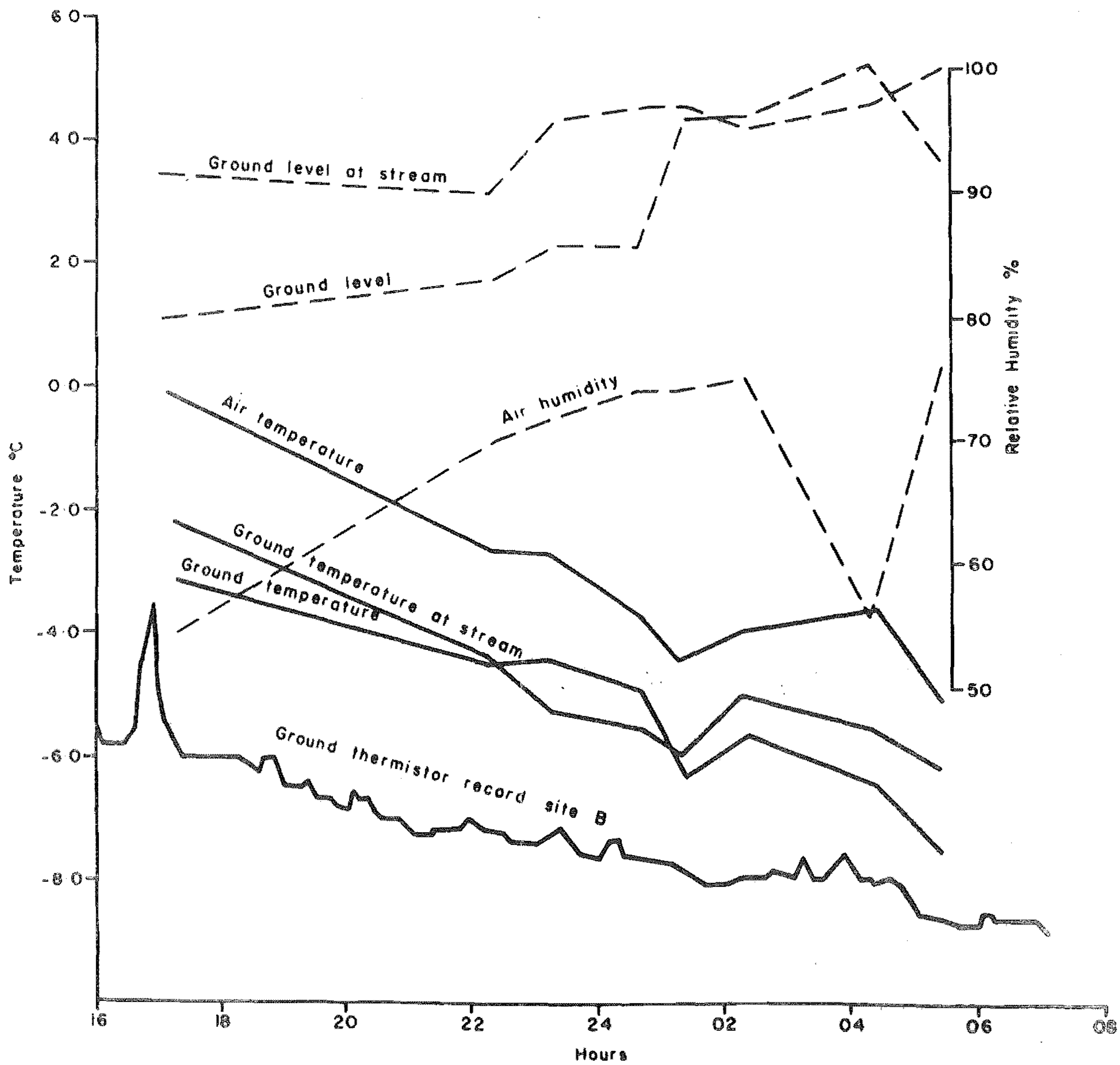


FIGURE 3.11: Values of air temperature and humidity during ice needle growth on July 8-9, 1967

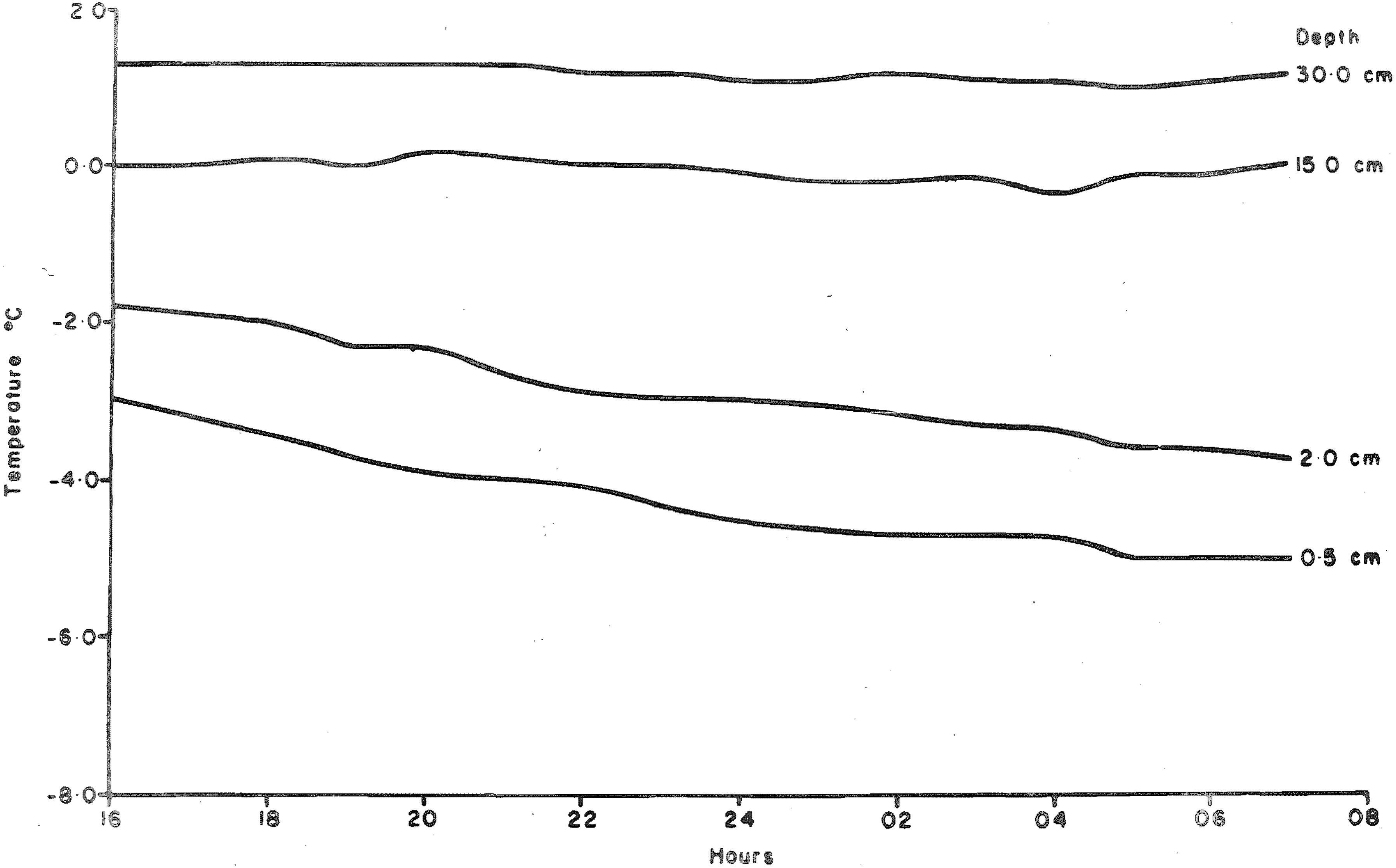


FIGURE 3.12: Values of soil temperatures during ice needle growth on July 8-9, 1967

periods were definitely underestimated, mainly due to the lysimeters not being placed completely within the area of directly applied heat. Sensible heat loss during the heating periods was calculated using a formula presented by Fishenden and Saunders (1953).

Ice needles similar to those observed in the field grew in the laboratory. Details of their growth, temperatures, and soil moisture conditions are shown in Fig. 3.13. The soil moisture records should be approached with caution since the gypsum blocks used were temperature dependent. Two significant effects of the repetitive freeze-thaw cycles may be noted. First, the ice needles, although growing in each cycle to about 3 cm, became progressively less dense after the second cycle. Second, thickening and consolidation of a layer of frozen soil at the base of the needles also occurred after the second cycle. Furthermore, the structure of the soil changed from being in a 'smooth' and well packed condition to becoming highly broken and covered with small, loose soil pellets.

While a much more detailed description of this work has been reported by Soons and Greenland (1970), of direct concern here is the heat balance of the surface during these events. Definition of the outer effective surface for which heat balance calculations are made is necessary, and here an important difficulty arises. The surface changes through time both in position, since the soil is uplifted, and in nature, since the ice needles present a much greater surface area than does the soil. Such changes render impossible direct instrumental measurements of heat balance components

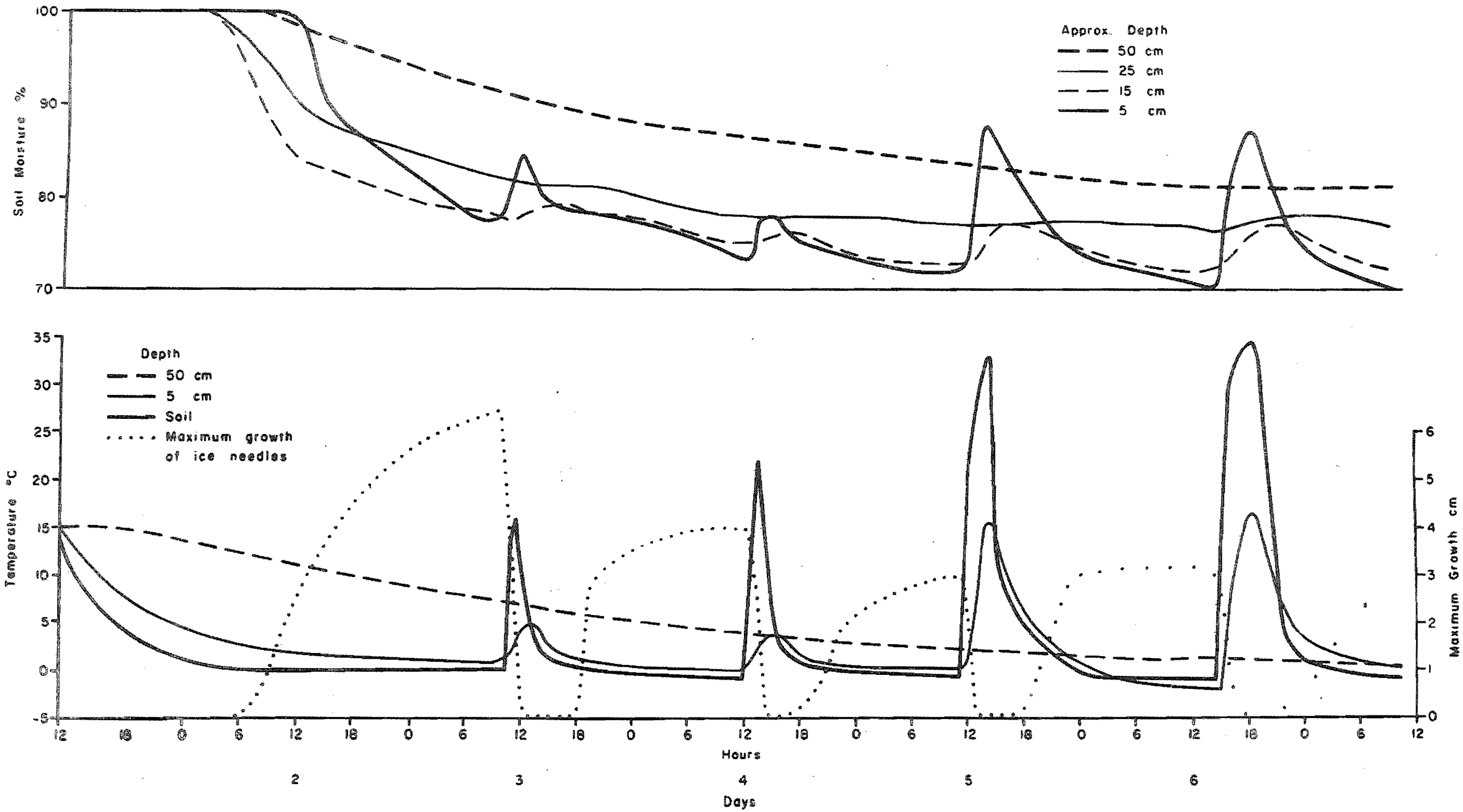


FIGURE 3.13: Values of soil temperatures and percentage available soil moisture during laboratory growth of ice needles

on a real microscale, using available methods. The measurements described here still generalise the energy exchanges that occur to some extent. The limitations of measuring heat balance components in this case must also be borne in mind. During the freezing periods soil heat flow was measured 1.0 cm below the surface from which the needles began to grow, and evaporative heat loss was measured some 15.0 cm above this surface. It was therefore assumed that the outer effective surface included a layer of soil 1.0 cm deep and all the ice needles and the soil that they lifted. The same outer effective surface was assumed for the thawing periods.

Since the soil surface was at approximately the same temperature as the air and surrounding walls of the room during the freezing periods, a near equilibrium of radiative heat exchange was assumed. During the freezing period the most significant heat flux was upward through the soil. Because of the temperature similarity P would be small. Most of the heat arriving at the surface was therefore lost in evaporation. The heat balance for this period therefore resolves itself to a gain of heat by soil heat flow and a loss by evaporation (sublimation). The data of Table 3.7 show further, that during the second, third and fourth periods LE was greater than could be supported by A. This points to the importance of the heat released when water changes to ice.

The results for the thawing periods (Table 3.8) are reasonable insofar as the partitioning of available energy is concerned. LE is the principal heat sink, but P and A are both significant. Only the first two thawing periods bear any resemblance to the field situation since Rn values for the last two periods are up to ten times higher than could be found in the field. An interesting feature shown by the continuous records of A during thawing, was that heat passed into the soil at a fairly high rate initially, but the flow was steadily reduced during the remainder of the period. This decrease may be due to the reduction of the thermal conductivity of the soil as first the proportion of ice, and then the proportion of water, is reduced.

A comparison of the heat balances in the field and the laboratory is seen in Table 3.9. In the field A was not measured directly under the ice needles but about 3 m away. P was assumed to be negligible. Clear sky conditions with consequently low LW$\downarrow$ help to explain the negative Rn value. A gain of heat at the surface, shown in the balance column (Table 3.9) for July 8, may represent heat released in the formation of ice from soil water and/or in the condensation (sublimation) of water vapour onto the cold surface.

It can be seen that values of A in the field are considerably higher than those in the laboratory. The total soil heat flow for the period 1600 hr July 8 to 0700 hr July 9, is 61.5 ly which is comparable to 58.0 ly lost during the 47 hrs of the first laboratory freezing period. However, the total flow for the same hours on June 28 is 105 ly, while the average soil heat flow for the last four laboratory freezing periods is only 17 ly. Estimates of temperature gradients in the top 5 cm show similar differences; those for the laboratory averaging about 0.2^{o}C cm^{-1} and that

TABLE 3.7

SUMMARY OF HEAT BALANCE FOR FREEZING PERIODS. UNITS LY

Period	Total Soil Heat Flow	Evaporative Heat Loss	Balance of Heat
1	+ 58.0	- 49.4	- 8.6
2	+ 12.2	- 17.6	+ 5.4
3	+ 10.7	- 18.0	+ 7.3
4	+ 22.0	- 28.0	+ 6.0
5	+ 23.9	- 19.9	- 4.0

TABLE 3.8

SUMMARY OF HEAT BALANCE FOR THAWING PERIODS. UNITS LY

Period	Evaporative Heat Loss	Convected Heat Loss	Soil Heat Flow	Resultant Net Radiation
1	- 62.0	- 10.8	- 22.5	+ 95.3
2	- 67.0	- 15.3	- 24.0	+ 106.3
3	-151.0	- 57.2	- 54.0	+ 262.2
4	-202.0	-76.0	- 72.0	+ 350.0

TABLE 3.9

HEAT BALANCE IN THE FIELD AND LABORATORY DURING FREEZING PERIODS. UNITS LY HR^{-1}

Period	Soil Heat Flow	Evaporative Heat Loss or Balance	Convective Heat Loss	Net Radiation
Field:				
1600 June 28 to 0700 June 29	+ 7.0	- 1.0	-	- 6.0
1600 July 8 to 0700 July 9	+ 4.1	+ 3.1	-	- 7.2
Laboratory:				
1	+ 1.2	- 1.1	-	-
2	+ 0.5	- 0.7	-	-
3	+ 0.5	- 0.8	-	-
4	+ 0.9	- 1.2	-	-
5	+ 1.5	- 1.3	-	-

for the field (July 8) being 1.7^{o}C cm^{-1}. Since ice needles grew in all cases, it appears that there is a large range of rates and totals of soil heat flow that can accompany ice needle growth. As long as the soil heat flow is directed towards the surface, the size and rate of flow is not necessarily critical, at least within the ranges quoted above.

The other major difference between the field and the laboratory heat balances, is the manner by which heat is lost from the surface. In the field the principal heat loss is by radiation while in the laboratory evaporation/sublimation is more important. It is possible that more heat is lost by evaporation in the field, and less is lost in the laboratory, where finite but small sensible and radiative heat losses probably occurred. In general however, the values in Table 3.9 present a clear picture of the heat flows in the two situations. Therefore, as far as the ice needle growth is concerned, the method by which heat is lost does not seem to be important.

Measurements of environmental conditions during needle ice growth in the field have also been made by Outcalt (1969, 1970) in British Columbia and Virginia. He found that factors affecting the morphology of frozen soils included the temperature gradient between the surface and the freezing plane and the degree of evaporation induced desiccation in melting periods. In the present work, with regard to the first factor, the temperature gradient between the surface and the freezing plane was not measured directly but the temperature gradients quoted above include the layer between the surface and the freezing plane. The results from the laboratory experiments show that temperature gradients in the top 5 cm can be quite small and still give rise to ice needle growth, other conditions being favourable. Outcalt (1969) has demonstrated that if the gradient and soil heat flow to the surface were too large, then segregation would cease, and the soil would be 'normally frozen' i.e. without ice needles. The field observations on 28 June show that A can be as large as 7 ly hr^{-1} and still give rise to segregation, but qualitative observations in the Chilton Valley, on other occasions, suggest that large temperature gradients are in fact accompanied by 'normally frozen' soil.

In the laboratory experiment desiccation during the melting periods certainly appeared to affect the morphology of the ice needles, as Outcalt suggests. The ice needles became progressively less dense after the second freeze-thaw cycle, until in the last cycle, there were only five or six stunted needles over the whole soil surface.

The present investigations show two other areas of agreement with Outcalt's observations. These are the rate of growth of needles during one cycle and the Rn value during growth. A rate of growth which is fast at the beginning and slow at the later stages (e.g. 0.8 cm hr^{-1} to 0.1 cm hr^{-1} in the first laboratory cycle) was also found by Outcalt. Furthermore, during growth, an average Rn value of about -6.0 ly hr^{-1} recorded in Virginia is very similar to the values of -6.0 and -7.2 ly hr^{-1} recorded in the Chilton Valley.

3.7 Summary

The most important points arising from this study of soil heat flow and related phenomena are as follows:-

1. Values of soil heat flow at 1 cm in the Chilton Valley show large variations on a day to day scale. On a seasonal scale an approximate balance is reached between heat flowing into the soil in winter and out of the soil in summer. In the study year there is a relatively fast transition between outflow from, and inflow to, the soil in the late spring, but a slow reverse transition in late summer and autumn. Soil heat flow, especially during the summer months, has a close relationship with values of net radiation.

2. Soil temperatures also show considerable day to day variability although this is reduced at the 15 cm and 30 cm levels. On the annual scale soil temperatures show evidence of the expected trends. During the study period maximum soil temperatures at all depths occurred two months after the maximum radiation amounts, although this lag is rather confused since Rn values showed a double maxima in this period. There is evidence of the occurrence of freeze-thaw cycles at the recorder site on more than one third of the days in May, June and July. Monthly mean soil temperatures at 2 cm and 15 cm suggest that the soil heat flow totals measured by the flux plates may be underestimates.

3. Many thermal characteristics of the soil show marked differences between examples of days taken in different seasons. Soil temperatures in a winter example were affected markedly by an input of rain water, and had higher values despite a decreased value of Rn. In spring and summer examples, input of rain water and lower Rn values were associated with lower soil temperatures and decreased diurnal ranges. The thermal conductivity and heat capacity of the soil are related to soil moisture conditions, and the value of the former show a marked difference between the spring and summer examples and the winter example.

4. Laboratory and field observation of ice needle growth suggests that there is a range of at least 0.5 and 7.0 ly hr^{-1} in the value of A under which ice needle growth is possible. Rn values of -6 and -7 ly hr^{-1} have been found during ice needle growth, but the actual method of heat loss appears to be of little significance.

CHAPTER FOUR

EVAPOTRANSPIRATION AND EVAPORATIVE HEAT LOSS

4.1 Introduction

The process of evapotranspiration requires heat energy, and therefore represents a heat loss at the surface of the earth. Evapotranspiration is directly and indirectly associated with plant growth, and through this, affects surface cover and geomorphological processes. Evaporation from bare surfaces can also influence the rate of some land sculpturing processes. An assessment of the quantity of water lost to the air is often an integral part of studies in hydrology. The magnitude of evaporative heat loss is therefore important for biological, geomorphological and hydrological reasons.

Many methods exist for measuring or estimating rates of evapotranspiration and the latent heat flux (LE). 'That so many exist indicates that none is completely adequate for all purposes' (Sellers, 1965, p.156). In the present study evapotranspiration (ET) was measured for long time periods by non-weighing lysimeters. In addition, over a period of a month in the summer, evaporation (E) was measured from an open pan and potential evapotranspiration (PET) was estimated by means of empirical formulae. It was also possible to use the lysimeters to estimate actual evapotranspiration (AET) for a period of sixteen days during this month. Examination of all of these data made it possible to develop a method of obtaining daily values of AET.

Definition of the terms potential and actual evapotranspiration has given rise to much debate. PET has been defined as the 'maximum quantity of water capable of being lost, as water vapour, in a given climate, by a continuous stretch of vegetation covering the whole ground when the soil is kept saturated. It thus includes evaporation from the soil and transpiration from the vegetation from a specified region in a given time interval' (W.M.O., 1966). Scientists of the World Meteorological Organisation (1966) have also defined AET as the 'sum of the quantities of water vapour evaporated by the soil and transpired by plants under existing meteorological and soil moisture conditions.' Thornthwaite (1944) defined PET as 'the water loss from a moist soil trace completely covered by vegetation, and large enough for oasis effects to be negligible.' A more specific definition of PET was given by Penman (1956). Instead of the term potential evapotranspiration Penman prefers to use potential transpiration which he defines as 'the amount of water transpired in unit time by a short green crop, completely shading the ground, of uniform height and never short of water'. Some of the controversy appearing in the literature on PET would be resolved if more consideration were given to the fact that the empirical formulae of Thornthwaite (1948) and Penman (1948, 1956) for estimating PET, refer to each author's own definition of the parameter. In the present study, although the empirical formulae of Thornthwaite and Penman are used, an attempt is made to measure or estimate AET and PET as defined in the above statements of the World Meteorological Organisation.

In this chapter measurements made with the lysimeters are discussed, and an examination of ET during part of the summer is described. The lysimeter measurements and conclusions from data collected in the summer form the basis for the development of a method for obtaining daily values of AET. It is then possible to present the values of AET and LE that occurred in the study year. Finally, studies of the phenomena of soil melting and drying are reported. These phenomena are related to ET. Also, to a certain extent, data from these studies were helpful in the development of the method of obtaining daily AET values.

4.2 Lysimeter Measurements of Evapotranspiration

AET was measured in the field by four non-weighing lysimeters or evapotranspirometers. These followed the Thornthwaite model as modified by Mather (1954) and have been used elsewhere when more elaborate weighing lysimeters have been unavailable and/or impracticable (Green, 1958; W.M.O. 1966). The field tank consisted of a 44 gallon drum with a surface area of 2370 cm^2. Gravel was placed at the bottom of the tank and the soil profile was 'reconstructed' on top of this, giving a depth of soil of between 30-35 cm. A percolation receiving tank was situated down slope from the field tank. Four lysimeters were installed in a line across the valley near the main recording site (Fig 1.3 and Plate 3). Plates 4-8 show the lysimeters and

PLATE 3: Positions of Lysimeters (marked by arrows)

PLATE 4: Lysimeter 2

PLATE 5: Vegetation of Lysimeter 1

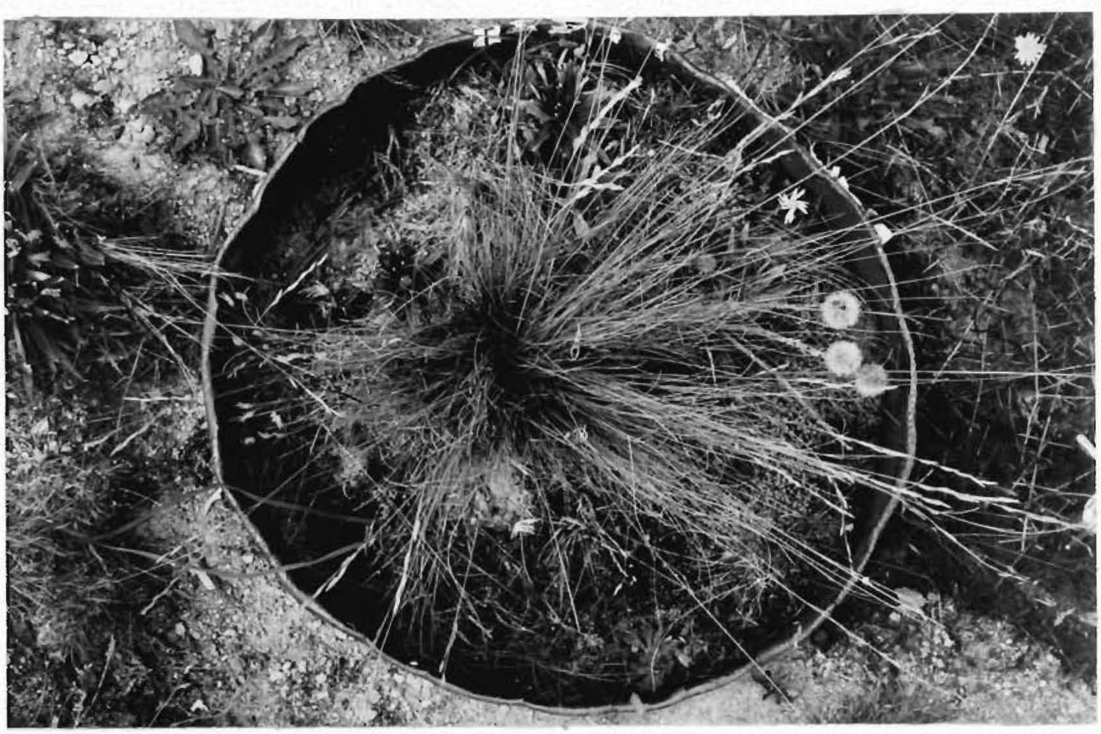


PLATE 6: Vegetation of Lysimeter 2

PLATE 7: Vegetation of Lysimeter 3

PLATE 8: Vegetation of Lysimeter 4

the different surface vegetation cover in each. It can be seen that the cover was not homogeneous for the four tanks. The tussock in lysimeter 4 actually dominates its field tank far more than the tussocks in the other field tanks, all of which contain other smaller plants such as the mountain daisy (Celmisia spectabilis). A rain gauge of 15.2 cm diameter was placed next to each field tank, with the orifice, as far as was possible, at the same level as the tank itself.

The tanks were placed in position in December 1968 and were allowed nine months for the vegetation and soil to settle down. Readings were then taken approximately every two weeks. Since no continuous record of soil moisture was kept, the results from the lysimeters combine this with the AET term of the water balance equation (Appendix G). Therefore, except for special circumstances (see section 4.3), the lysimeters cannot be used for short period estimates of AET. Over the annual period however, it may be satisfactory to assume no change in the value of stored soil moisture (Sellers,1965 p.82). Both soil moisture measurements by means of gypsum blocks, and the Thornthwaite method (Thornthwaite and Mather,1957) of computing the soil moisture term (Fig. 4.14), are in agreement with this assumption.

Measurements of AET using the type of lysimeters described above are subject to possible errors. These errors are discussed fully in Appendix G. Random errors arise from the measurement of the surface catchment area of the field tanks and lysimeters. These lead to an error of approximately 1% of the annual total AET (Appendix G). Other errors are considered to be partly or wholly systematic. These are due to the following factors:-

1. The absence of a runoff term in the water balance equation of the field tank.
2. The different sizes of the raingauge and field tank catchment areas.
3. Interception of rainfall by protruding vegetation in the field tank.
4. In-splash into rain gauges and field tanks.
5. The possibility of the field tank prohibiting the free flow of water in the soil.

In Appendix G, it is shown that some of the errors due to these factors work in opposite directions, and/or are small. Errors due to 1 and 2 are small. Errors due to 3 and 4 will probably lead to underestimates of AET whereas errors due to 5 will lead to an overestimation.

The results from the lysimeter measurements (Table 4.1) show that the AET for the whole year, averaged over the four lysimeters, was 56 cm. This is 0.51 of the precipitation (111 cm). Despite the inhomogeneity of the surface cover and situation of the lysimeters, there is little difference between the four sets of values. A non-parametric statistical test, the Kruskal-Wallis test (Blalock,1960 p.265), showed both the rainfall and percolation values to be drawn from the same continuous population (significant at the 0.001 level). The short period values show that the ratio of percolation to rain water was highest in the winter and early spring and lowest in the summer and autumn.

Rain water that is not lost in percolation is used to recharge soil moisture or in ET. The values of the

TABLE 4.1

MEASUREMENTS FROM LYSIMETERS IN CMS

Date	Lysimeter 1			Lysimeter 2			Lysimeter 3			Lysimeter 4		
	Rain	Perc.	Remainder	Rain	Perc.	Remainder	Rain	Perc.	Remainder	Rain	Perc.	Remainder
11.9.69.	14.7	10.2	4.5	15.9	10.9	5.0	14.7	10.7	4.0	14.7	10.0	4.7
25.9.69	8.6	6.7	1.9	8.2	6.7	1.5	8.1	7.5	0.6	8.1	6.4	1.7
4.10.69	0.0	0.0	0.0	0.0	0.0	0.0	0.1	0.0	0.1	0.2	0.0	0.2
23.10.69	2.5	0.0	2.5	2.6	0.2	2.4	2.7	0.0	2.7	2.6	0.0	2.6
5.11.69	3.3	1.7	1.6	3.3	1.0	2.3	3.4	1.7	1.7	3.4	1.4	2.0
19.11.69	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
3.12.69	3.9	0.1	3.8	3.9	0.4	3.5	3.9	0.1	3.8	3.9	0.2	3.7
16.12.69	1.9	0.0	1.9	2.3	0.0	2.3	2.1	0.0	2.1	2.2	0.0	2.2
28.12.69	17.2	10.5	6.7	18.1	10.0	8.1	16.9	10.1	6.8	16.5	9.8	6.7
12.1.70	2.9	0.0	2.9	3.0	0.3	2.7	2.9	0.1	2.8	2.7	0.1	2.6
20.1.70	2.1	0.0	2.1	2.3	0.0	2.3	2.0	0.0	2.0	2.6	0.0	2.6
24.1.70	7.3	4.6	2.7	7.3	4.2	3.1	7.3	3.9	3.4	8.0	4.4	3.6
30.1.70	0.7	0.0	0.7	0.6	0.0	0.6	0.7	0.0	0.7	0.8	0.0	0.8
4.2.70	0.8	0.0	0.8	0.7	0.0	0.7	0.7	0.0	0.7	0.7	0.0	0.7
9.2.70	2.4	0.0	2.4	2.4	0.1	2.3	2.4	0.0	2.4	2.4	0.1	2.3
16.2.70	0.2	0.0	0.2	0.4	0.0	0.4	0.4	0.0	0.4	0.3	0.0	0.3
26.2.70	0.3	0.0	0.3	0.4	0.0	0.4	0.6	0.0	0.6	0.5	0.0	0.5
11.3.70	5.2	0.7	4.5	5.3	0.6	4.7	5.1	0.7	4.4	5.2	0.9	4.3
25.3.70	2.1	0.0	2.1	2.3	0.0	2.3	2.0	0.0	2.0	2.0	0.0	2.0
9.4.70	4.3	1.5	2.8	4.4	0.7	3.7	4.1	1.2	2.9	4.5	1.6	2.9
23.4.70	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0
8.5.70	0.8	0.0	0.8	1.3	0.0	1.3	0.8	0.0	0.8	0.8	0.0	0.8
4.6.70	8.0	4.6	3.4	8.0	4.2	3.8	8.0	4.0	4.0	8.4	4.6	3.8
17.6.70	4.2	2.7	1.5	5.0	3.0	2.0	3.9	3.1	0.8	4.1	3.0	1.1
1.7.70	1.2	1.5	-0.3	1.0	1.1	-0.1	1.1	0.8	0.3	1.1	1.0	0.1
18.8.70	15.2	10.9	4.3	15.6	10.9	4.7	15.0	11.2	3.8	15.3	11.2	4.1
TOTALS	109.8	55.7	54.1	114.3	54.3	60.0	108.9	55.1	53.8	111.0	54.7	56.3

combination of these latter water uses are seen (Table 4.2, Fig. 4.1) to have been higher in summer, as might have been expected. However, there are periods in September/October, November, February, April and June (i.e. all seasons) when water use was less than 0.5 mm day^{-1}.

The above values may be compared with those of other workers. Rowley (1970), used almost identical lysimeters to investigate AET of narrow leaved snow tussock grassland at 915 m on the Rock and Pillar Range in Central Otago. During the period November 1966 to October 1967 she found AET values of 33.8, 42.9, 51.9 and 54.6 cm for snow tussock, burnt snow tussock, clipped snow tussock and blue tussock respectively. Her evidence indicated that the AET values were equivalent to PET values in this period. The Thornthwaite PET estimate for the same period was 49.1 cm and there was 99.6 cm of precipitation. These values are in the same order as those obtained in the Chilton Valley, but there are three important differences in the two investigations that should be noted. Firstly, the Chilton Valley lysimeters were occupied by lower sub-alpine flora other than tussock. Secondly, Rowley noted high interception values during some periods and these are not believed to have occurred in the present study (see Appendix G). Thirdly, her measurements indicated PET to have occurred throughout the year, while in the study period in the Chilton Valley, soil moisture deficits (see sections 4.5 and 4.6) caused AET to be less than PET at certain times (Fig. 4.6).

Archer and Collett (1970), working in the High Country near Mount Cook at 850 m also found evidence of AET values being lower than PET values, owing to soil moisture deficits

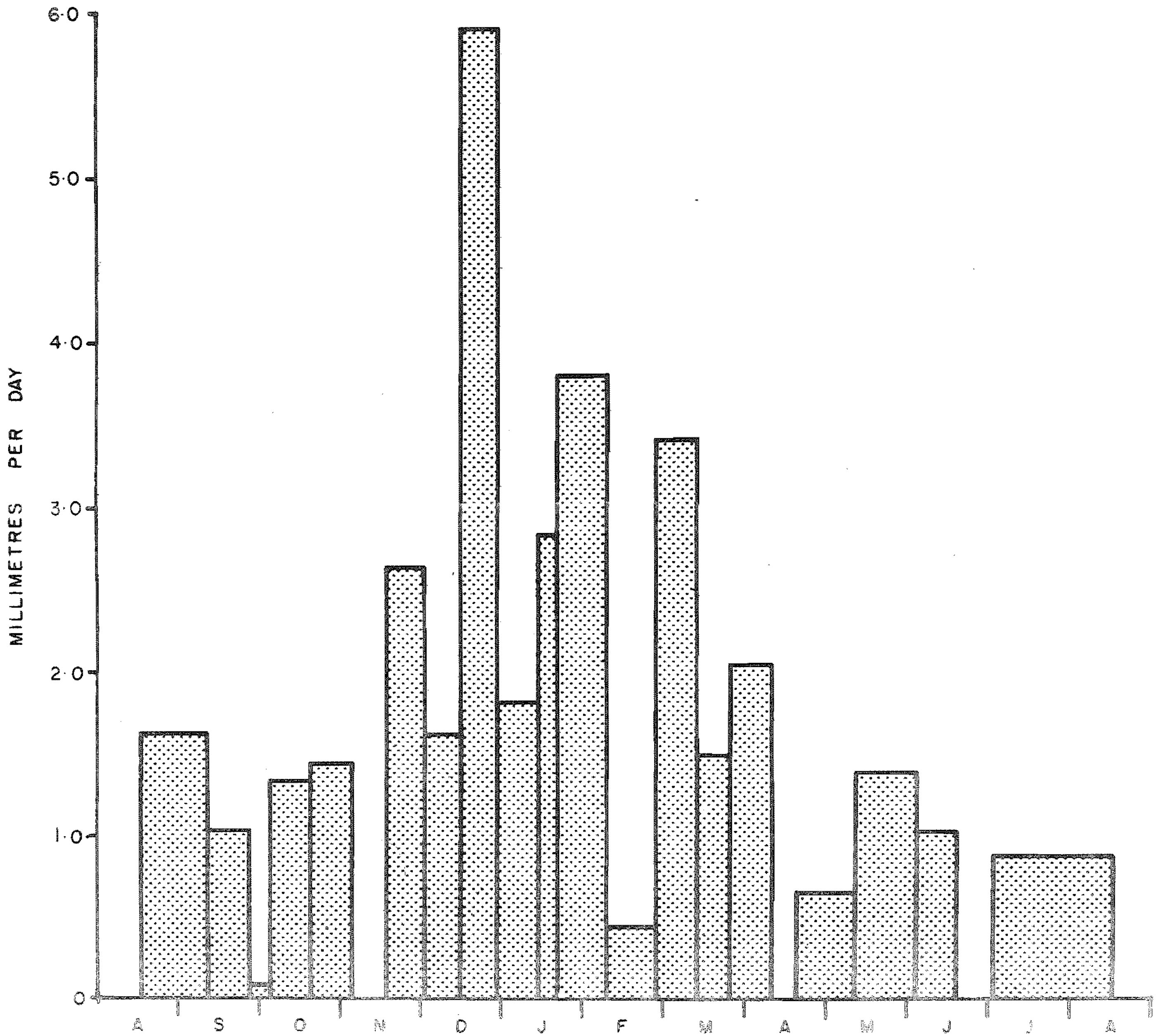


FIGURE 4.1: Values of water use per day measured by lysimeters

TABLE 4.2

AVERAGE WATER USE PER DAY IN mm DURING PERIODS ENDING ON THE DATES SHOWN

Date	Water Use
11.9.69	0.16
25.9.69	0.10
4.10.69	0.01
23.10.69	0.13
5.11.69	0.15
19.11.69	0.00
3.12.69	0.26
16.12 69	0.16
28.12.69	0.59
12.1.70	0.18
20.1.70	0.28
24.1.70	0.80
30.1.70	0.12
4.2.70	0.15
9.2.70	0.47
16.2.70	0.05
26.2.70	0.05
11.3.70	0.34
25.3.70	0.15
9.4.70	0.21
23.4.70	0.0
8.5.70	0.06
4.6.70	0.14
17.6.70	0.10
1.7.70	0.00
18.8.70	0.09

Average Remainder = 56.05 cms
560.5 mm

in summer. For the period October 1968 to December 1969, when 176.2 cm of precipitation was recorded, water loss from a raised pan evaporimeter was 110.9 cm. These authors reduced this figure by a factor of 0.75 to obtain a PET value of 80.6 cm. This reduction factor may be too high, judging by values for New Zealand given by Finkelstein (1961). At Craigieburn, at an altitude of 807 m, PET values based on pan evaporation, and Finkelstein's reduction factors, also indicate the possibility of soil moisture deficits in some years (Morris, 1965).

While it is dangerous to attach too much importance to E values from pans, in areas where wind velocities can be high, a general pattern begins to emerge. Although there appear to be no soil moisture deficits in the higher altitude Rock and Pillar area, some may occur in the slightly lower altitude mountain stations in Canterbury. A similar view, but solely in terms of altitude, has been expressed by Mark (1965), with reference to Otago.

4.3 Daily Evapotranspiration in a Summer Period

Daily measurements of E from a raised pan evaporimeter were made between January 20 and February 18, 1970. Additional measurements were taken in order to permit the computation of PET using empirical formulae. The method presented by Thornthwaite and Mather (1957) was used for the Thornthwaite calculations and the formula given by Penman (1956) was also applied.

Examination of the daily values (Table 4.3 and Fig. 4.2) show inertia effects to be present in the records of the pan

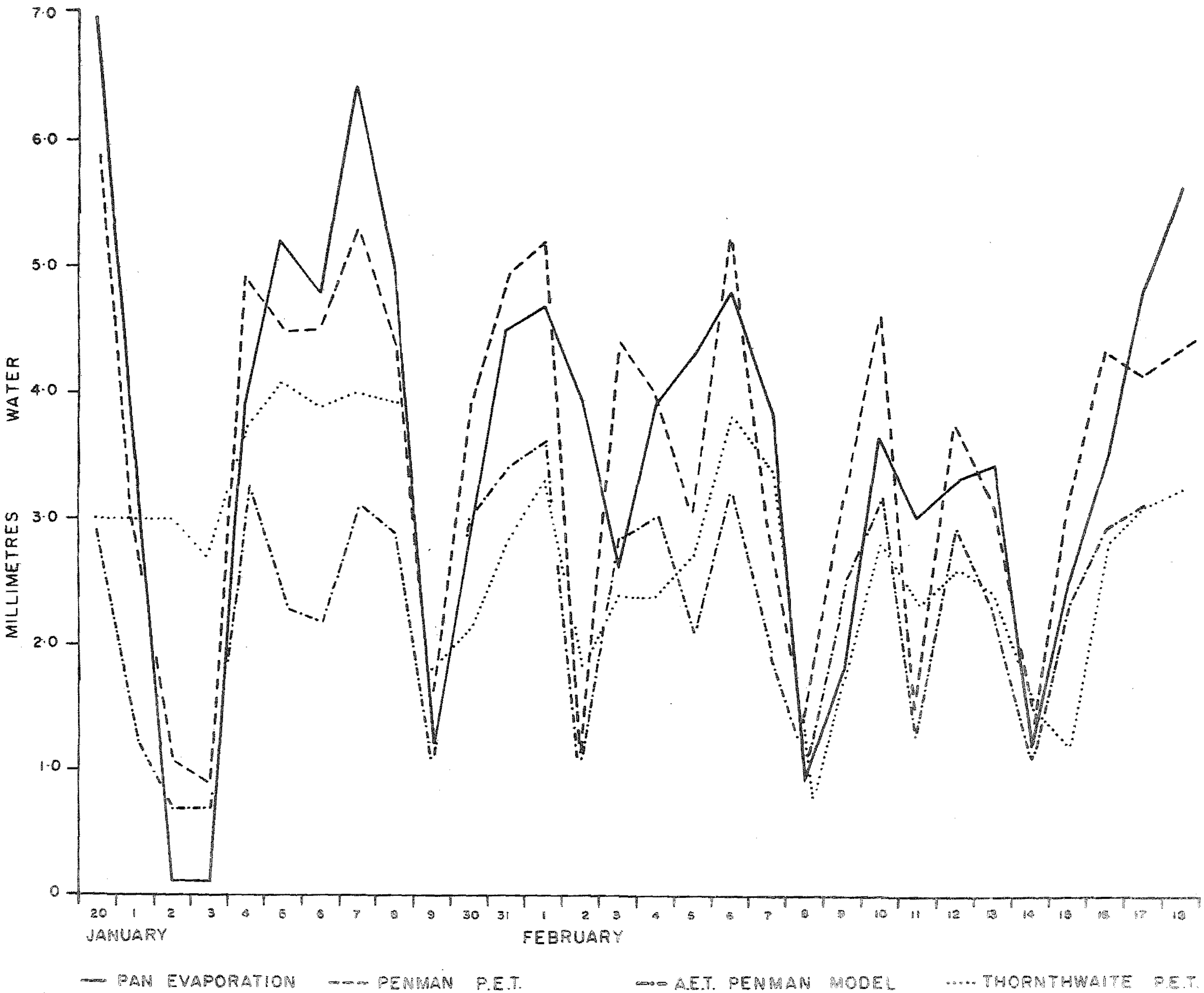


FIGURE 4.2: Daily values of E, AET, and PET for the period 20 January to 18 February 1970

TABLE 4.3

EVAPORATION AND POTENTIAL EVAPOTRANSPIRATION VALUES FOR JAN. 20 - FEB 18, 1970 AT CHILTON VALLEY RECORDER SITE

Date	Pan Evaporation mm	P.E.T. Thornthwaite mm	P.E.T. Penman mm	A.E.T. Modified Penman Model mm
Jan. 20	7.0	3.0	5.9	2.9
21	3.5	3.0	3.0	1.3
22	0.1	3.0	1.1	0.7
23	0.1	2.7	0.9	0.7
24	3.9	3.7	4.9	3.3
25	5.2	4.1	4.5	2.3
26	4.8	3.9	4.5	2.2
27	6.4	4.0	5.3	3.1
28	5.0	3.9	4.4	2.9
29	1.2	1.8	1.6	1.1
30	2.9	2.1	3.9	3.0
31	4.5	2.8	4.9	3.4
Feb. 1	4.7	3.3	5.2	3.6
2	3.9	1.8	1.1	1.1
3	2.6	2.4	4.4	2.8
4	3.9	2.4	4.0	3.0
5	4.3	2.7	3.0	2.1
6	4.8	3.8	5.2	3.2
7	3.8	3.4	2.9	1.9
8	0.9	0.8	1.4	1.0
9	1.8	1.7	3.1	2.5
10	3.6	2.8	4.6	3.2
11	3.0	2.3	1.5	1.3
12	3.3	2.6	3.7	2.9
13	3.4	2.4	3.1	2.3
14	1.2	1.4	1.2	1.1
15	2.5	1.2	3.0	2.3
16	3.5	2.8	4.3	2.9
17	4.8	3.1	4.1	3.1
18	5.6	3.2	4.4	3.2
Total Jan.23-Feb. 8	62.9	49.6	62.1	40.7
Jan.20-Feb.18	106.2	82.1	105.1	70.4

evaporimeter on the days following 15 February, for example, the pan values continue to increase when the other methods show only small increases or a decrease. Also, on February 3, pan values continue to decrease when the other methods indicate an increase in the amount of water loss. The Penman method frequently gives the highest values of PET. On 13 out of the 30 days, the resulting Penman values exceed the water equivalent of Rn. Most of these days are associated with relatively high saturation vapour pressure deficits (4-8 mm Hg), and/or high wind run totals (100 - 220 miles day^{-1}). Despite the high values of PET indicated, the Penman method appears to be more sensitive to daily weather changes than the Thornthwaite method. The latter also seems to show some inertia or lag effects, as, for example, in the period 20-24 January. Inertia is commonly found in annual records given by this method, owing to the lag of air temperature changes behind radiation changes (Sibbons, 1962). It is reasonable that such inertia may also occur on a daily basis. While inertia effects may not be serious for long term estimates of PET, they may have an adverse effect on daily estimates.

It is possible to estimate the total PET for the period 24 January to 8 February by means of the lysimeter values. Rain, on and preceding January 23, brought the soil of the field tanks to field capacity. This is shown by the high soil moisture values (Fig. 4.12), and the fact that percolation occurred. These conditions were not observed again until February 8, when there was only a trace of percolation. Two light rainfalls occurred on January 30 and

February 3, but gave no percolation. Since the soil moisture records showed continuously high values throughout this period, except for the last two days, it may be assumed that PET occurred for the whole period with the possible exception of the last two days. Therefore the total PET may be approximated by the total rainfall from January 24 up to, and including, February 8 since all of this was used to recharge soil moisture lost by ET. Neglecting the very small amount (less than 0.1 mm) of percolation on February 8, and allowing for AET values to be less than PET values on the last two days, the total PET for the whole period was slightly greater than 38.0 mm, the total incoming rainfall for the period. Comparing this with the totals given by the other methods (Table 4.3), it represents 0.59 of the pan evaporation, 0.60 of the Penman estimate and 0.75 of the Thornthwaite estimate. The two latter values indicate that the empirical formulae give overestimates of PET for this period, and that the Penman method overestimates more than the Thornthwaite method.

An examination of the totals of PET for the whole month may be made. Finkelstein (1961) has suggested a factor of 0.55, for summer in New Zealand, by which raised pan E values may be reduced to values of PET. If this factor is applied to the raised pan data, the PET for the month at the Chilton Valley becomes 89.0 mm. This represents 1.08 of the Thornthwaite estimate and 0.85 of the Penman estimate. On this basis, the Thornthwaite method could be considered satisfactory while the Penman method still overestimates.

The conclusions from the results and discussion in this section are that for the period studied:

1. The Thornthwaite method is the most satisfactory of the two empirical methods for giving totals of PET over a period of two weeks or more, but, owing to lag effects, is not entirely satisfactory for giving daily values.
2. The Penman method tends to give overestimates of values of PET, but appears more sensitive to daily weather changes.

4.4 A Model for Obtaining Daily Evapotranspiration Values

4.4.a. Introduction

An estimation of daily LE was necessary to fulfill the need for daily values of all heat balance components. The approach used in finding a method of obtaining daily estimates of AET is summarised in Fig. 4.3. As can be seen in this figure, a certain amount of data was available. It was possible to use the empirical formula of Thornthwaite to compute daily estimates of PET and AET. It was also possible, with the aid of certain assumptions, to use the Penman formula to obtain day to day values of PET. The results of all of these estimates were compared with the known data, and a decision was made as to which of the Thornthwaite or Penman methods to adjust further in order to give daily values of AET for the Chilton Valley. Adjustments to the method chosen were made with regard to the experience of other workers. Modifications were compatible with physical laws, the measured or computed values of the other heat balance components, and the available known data.

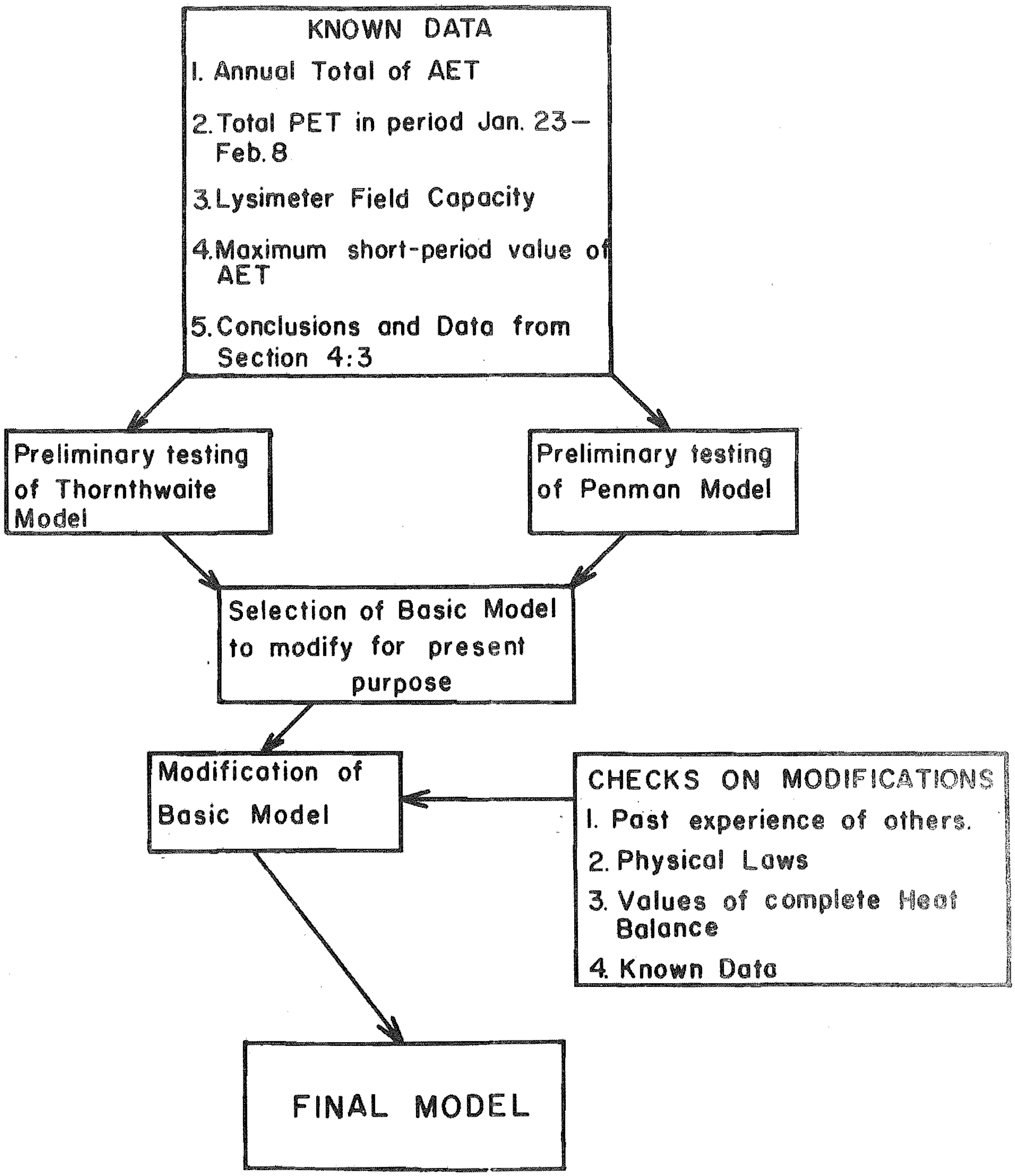


FIGURE 4.3: Diagram of approach used to obtain a method for estimating daily values of AET

4.4.b. Known Data

The known data consisted of the following:-

1. Lysimeter measurements showed that the value of AET for the study period was 56 cm.
2. Between January 23 and February 8, the value of PET was slightly more than 3.8 cm.
3. The field capacity of the lysimeters was found to be between 5.3 and 6.1 cm of water.
4. Since only the water available can be lost by ET, the field capacity of the field tanks can be used in conjunction with the lysimeter data, to determine a maximum value of AET in any two week period (see section 4.6).
5. The conclusions listed at the end of, and the data given in, section 4.3 can also be considered.

4.4.c. Preliminary Testing of the Thornthwaite Method

Attention was focused first on the use of the Thornthwaite method. The results of a computation of the monthly water balanc using this method (Thornthwaite and Mather, 1957) are shown in Table 4.4 and Fig. 4.4. To compute values of AET the appropriate tables for a fine, sandy loam, and a root depth of 50.0 cm, with a provisional water holding capacity of 75 mm, were used. These conditions represent the closest approximation, quoted by Thornthwaite, to those of the soil in the Chilton Valley. The analysis (Table 4.4) shows the total AET for the year to be 60.2 cm, which agrees quite well with the lysimeter measured total of 56.0 cm.

It is difficult to make a direct comparison of the Thornthwaite results with the lysimeter data in terms of the

TABLE 4.4

MONTHLY WATER BALANCE COMPUTED BY THE THORNTHWAITE METHOD.
WATER UNITS ARE EXPRESSED IN mm

	Aug.	Sept.	Oct.	Nov.	Dec.	Jan.	Feb.	March	Apr.	May	June	July	Aug.	Year
T °C	2.7	6.9	6.3	12.8	13.4	15.1	13.1	12.1	10.0	3.7	3.9	3.7	6.9	
I	0.39	1.63	1.42	4.15	4.43	5.33	4.30	3.81	2.86	0.63	0.69	0.63	1.63	
Unadj. PET	0.5	1.2	1.1	2.2	2.3	2.6	2.3	2.1	1.7	0.6	0.7	0.6	1.2	
Adj PET	13.6	36.0	39.4	79.5	87.3	101.0	74.2	67.5	47.0	15.1	15.8	14.8	32.8	624.0
Precip	119.0	184.9	54.2	35.9	175.1	98.2	46.2	59.9	33.3	77.3	43.5	12.2	80.3	909.1
Precip-PET	105.0	148.9	14.8	-43.6	87.8	-2.8	-28.0	-7.6	-13.7	62.2	27.7	-2.6	47.5	
Acc. Pot. Water Loss	-	-	-	-43.6	-	-2.8	-30.8	-38.4	-52.1	-	-	-2.6	-	
Soil Moist Storage *	75	75	75	40	75	72	49	44	36	75	75	72	75	
Change of Soil Moisture Storage	0	0	0	-35	+35	-3	-23	-5	-8	+39	0	-3	+3	
A E T	13.6	36.0	39.4	70.9	87.3	101.0	69.2	64.9	41.3	15.1	15.8	14.8	32.8	602.1
Deficit	0	0	0	8.6	0	0	5.0	2.6	5.7	0	0	0	0	
Surplus	105.0	148.9	14.8	0	87.8	0	0	0	0	62.2	27.7	0	47.5	

* A provisional water holding capacity for a fine sandy loam and root depth 50.0 cm of 75 mm has been used.

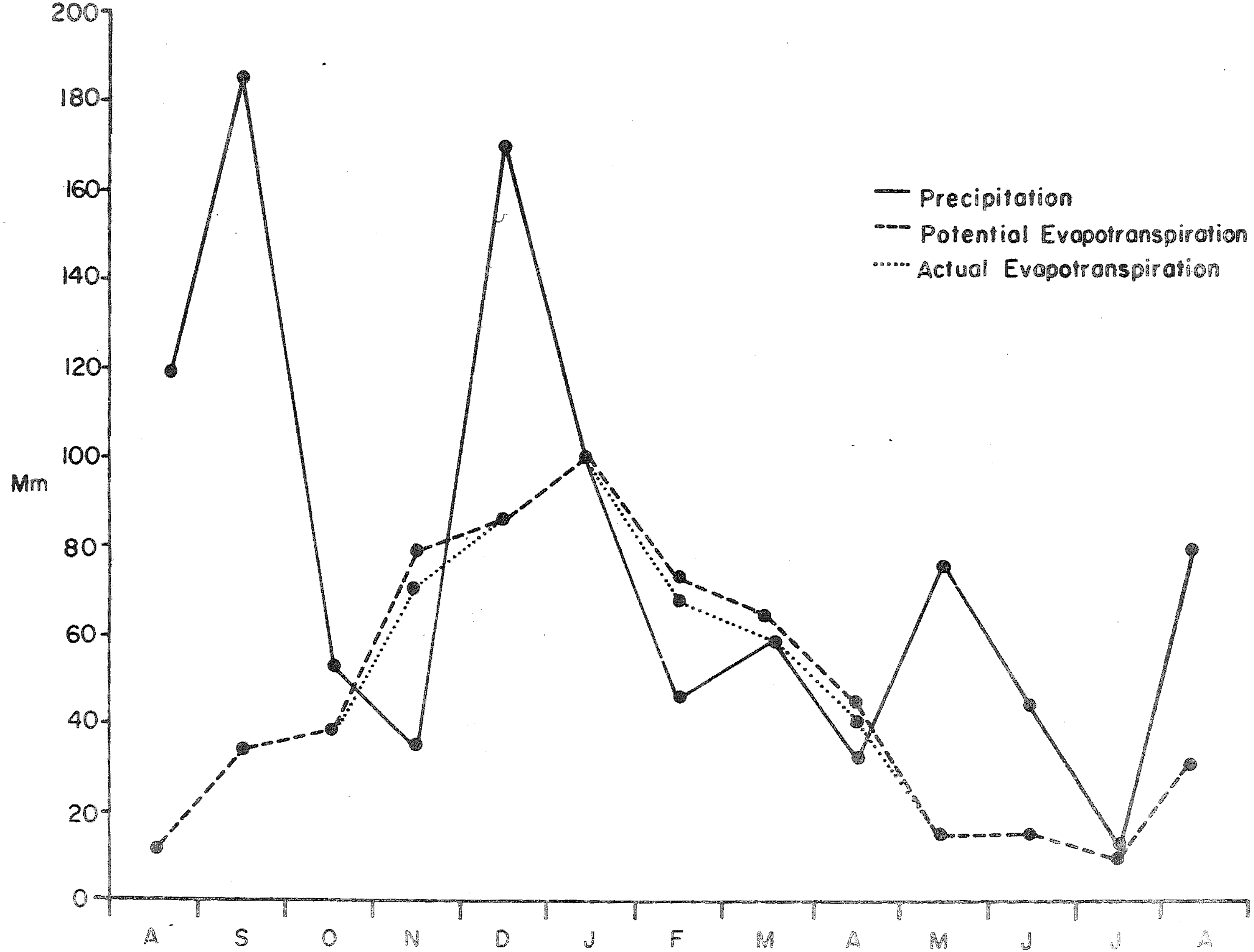


FIGURE 4.4: Monthly values of the water balance at the Chilton Valley during the study period, given by the method of Thornthwaite. (Thornthwaite and Mather, 1957)

value of soil moisture storage, since they are computed for a 50 cm soil column compared with the 30 cm column of the lysimeters. However, if the Thornthwaite AET amounts are applied to the lysimeter data, in a series of solutions of the water balance equation (as described in section 4.6 and shown in Table 4.10), the soil moisture term is negative for a large part of the last half of the year, and does not reach the value of field capacity at the end of the year. Since, in physical reality, the soil moisture term cannot be negative, the application of Thornthwaite AET values to the lysimeter data indicates that in some periods the Thornthwaite estimates demand the ET of more water than was actually available. This suggests that at least some of the Thornthwaite AET values are in fact overestimates. If the Thornthwaite estimates of AET are applied to a 50 cm soil depth with a capacity of 75 mm, soil storage reverts to field capacity in late May (Fig. 4.14). The depth of soil from which water can be drawn for ET may therefore be critical in the application of the Thornthwaite method. In testing the value of the Thornthwaite estimates of daily AET in this location, more weight must be given to the soil moisture results obtained using a 30 cm soil depth, rather than a 50 cm soil depth, because the actual root depth of vegetation in the Chilton Valley is nearer the lower value. The lower value of root depth also applies to other areas of the High Country. For example, Rowley (1970), who was studying a larger species of tussock than that dominant in the Chilton Valley, reports little penetration of roots below 35 cm. Therefore it may be concluded that the Thornthwaite estimates of daily AET, when tested by using the lysimeter

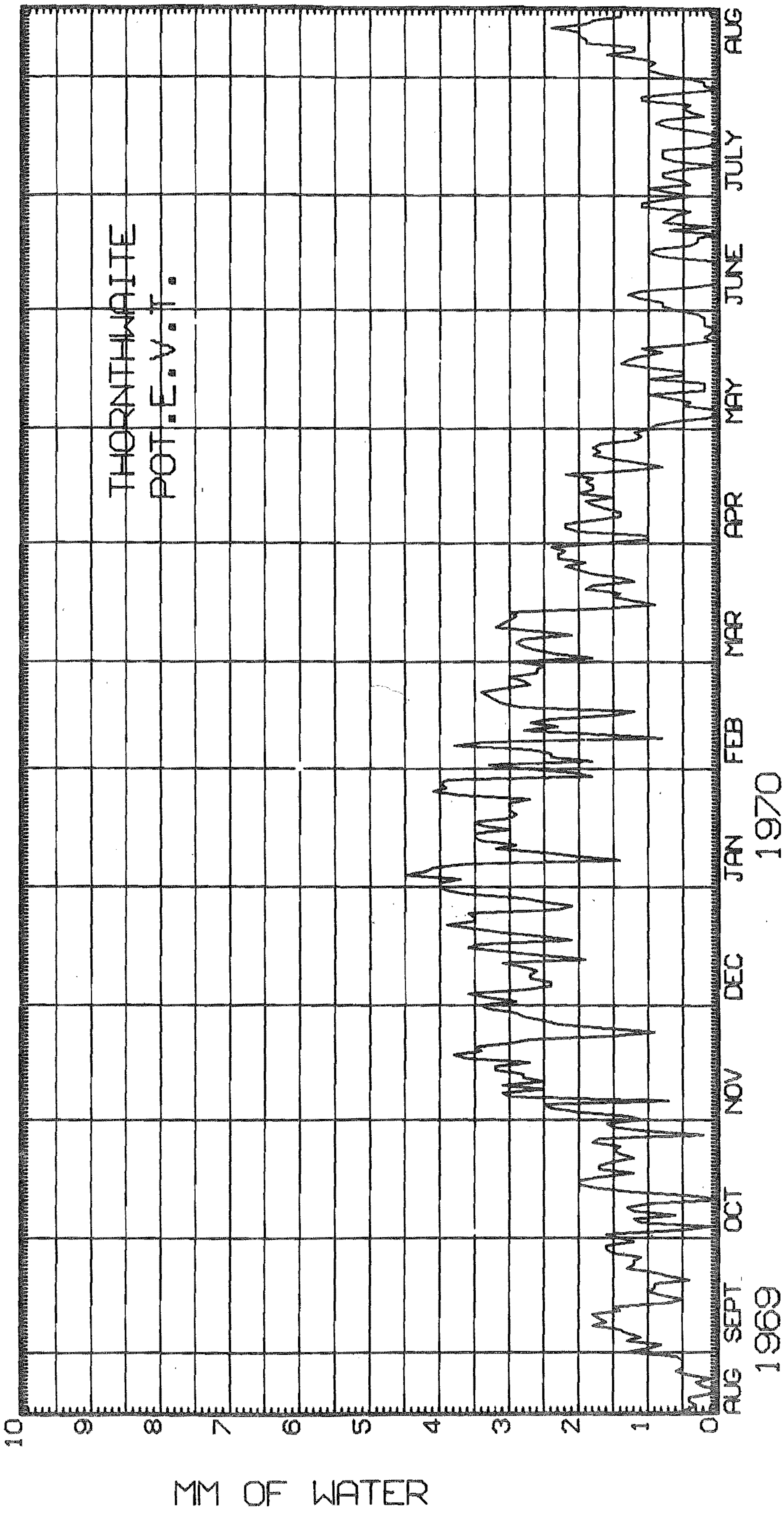


FIGURE 4.5: Daily values of PET at the Chilton Valley for the study period, given by the method of Thornthwaite. (Thornthwaite and Mather, 1957)

results, appear to give values that are too high, at least in some periods.

Two further aspects of the use of the Thornthwaite method are recalled. It was shown in the previous section that the PET value for January 24 to February 8 was only 0.75 of that indicated by the Thornthwaite method. It was further indicated that owing to a lag effect, daily values of PET estimated by the Thornthwaite method should be approached with caution. Daily Thornthwaite estimates of PET for the study period (Fig. 4.5) will be compared later with those given by a Penman method.

4.4.d. Preliminary Testing of the Penman Method

Penman's (1956) formula for estimating PET is

$$PET = \left(\frac{\Delta}{\gamma} Rn + Ea\right) \Big/ \left(\frac{\Delta}{\gamma} + 1\right) \text{ mm day}^{-1} \qquad 4.4.1.$$

where

$$Ea = 0.35 \quad (1 + u \times 10^{-2}) \quad (e_a - e_d) \qquad 4.4.2.$$

and is called the 'drying power' term.

Δ is the slope of the saturation vapour pressure curve at the ambient air temperature, γ is the psychrometric constant, u is the wind speed at 2m (miles day^{-1}), e_a is the saturation vapour pressure at mean air temperature (mm Hg) and e_d is the mean vapour pressure of the atmosphere (mm Hg). Penman (1963, p.41) has published some values of the dimensionless quantity Δ/γ which varies with temperature.

An initial model was prepared so that this formula could be applied on a daily basis throughout the year. Saturation deficit data were not available on a day to day basis, so assumptions concerning these, based on available daily data, were made. A simplification of the value of Δ / γ was also employed. The assumed values for these parameters are shown in Table 4.5. Data from the period January 20 to February 19, 1970 were classified into (1) rain-days, (2) days during which wind came either from the north west, or when a macroscale north west wind was channelled by the value to become a north east wind, and (3) other days. The saturation deficit values in Table 4.5 are the average values for each of these categories of days, and are obtained from data collected in the field during this period. The values of Δ / γ are taken from a graph of Δ / γ against temperature constructed from data in standard psychrometric tables (Marvin, 1941). The values agree with those of Penman (1963). Wind velocities at 3 m were adjusted to give values at 2 cm by means of the power law (equation 5.4.1.) with an exponent of 0.25.

The application of this model to the study year gives a value of PET of 97.8 cm. Since the Thornthwaite method indicated a difference of only 2.2 cm between PET and AET for the year, the Penman model almost certainly leads to an overestimation. This overestimation is also reflected in the computed soil moisture terms (Table 4.10). The soil moisture storage in the lysimeters becomes negative in October and then remains negative for the rest of the year. This Penman model gave a total of 6.4 cm of PET for the period January 24 to February 8, which is 1.7 times greater than the lysimeter estimate. However, the sensitivity of the basic Penman method to daily weather changes, which was demonstrated in section 4.3 where no assumed data were used, was normally maintained. This can be inferred when daily values of PET

TABLE 4.5

APPROXIMATIONS AND ASSUMPTIONS USED IN THE PENMAN COMPUTATIONS OF DAILY PET VALUES FOR THE STUDY PERIOD

1) Saturation Deficit (mm Hg) (Derivation described in text)

	Initial Model	Modified Model	Applicability
$e_a - e_d =$	1.33	0.80	When mean air temp $\leqslant 4.5^oC$
$e_a - e_d =$	1.69	1.01	On any day when rain fell.
$e_a - e_d =$	8.00	4.80	When the wind was between 310 and 80 degrees and no rain fell.
$e_a - e_d =$	6.70	4.02	On other days.

2) Δ / γ Slope of Vapour Pressure curve at mean air temperature divided by the psychrometric constant.

Air Temperature ^oC	Δ / γ
< 5.0	0.83
$\geqslant 5.0$ and < 11.0	1.11
$\geqslant 11.0$ and < 15.0	1.55
$\geqslant 15.0$ and < 17.5	1.85
$\geqslant 17.5$ and < 20.0	2.15
$\geqslant 20.0$	2.50

using the Thornthwaite method (Fig. 4.5) are compared with daily values from the final modified Penman model (Fig. 4.6). Further supporting evidence is given in section 4.4.g.

4.4.e. Selection of the Most Suitable Basic Method for the Present Study

It is clear that neither the Thornthwaite nor the Penman formulae, as used above, are entirely suitable to explain the observed values of AET for the Chilton Valley. In the past both formulae have been modified in such a way that they can give a better evaluation of PET in a particular place or area (e.g. Garnier, 1956; Tanner and Pelton, 1960). It is reasonable that a similar procedure should be followed here.

The choice of which of the two basic methods to modify is made on the basis of three criteria. Firstly, it must be considered, in quantitative terms, how well the estimates of the formulae compare with observed quantities of ET. In this respect the Thornthwaite method is considered the most satisfactory for both short and long term estimates. The Penman method gives relatively large estimates, of PET. However, there is a certain degree of consistency in this overestimation since it occurs for almost all of the year. In support of this, calculations from the data in Table 4.10 show that, except for the periods between 16 December 1969 to 30 January 1970 and 25 March to 8 April 1970, Penman estimates of PET exceeded Thornthwaite estimates of AET, by at least 0.7 mm, and at the most 57.9 mm, for all of the periods sampled. Since the difference between Thornthwaite estimates of PET and AET is only 22 mm for the whole year, it is probable that Penman estimates of PET exceeded Thornthwaite estimates of PET for most of the time.

A second criterion is that the physical basis of the formula must be considered. In this respect the Penman formula must be favoured. This formula takes into account more of the variables that are physically responsible for the loss of water to the air. Although some parts of the formula have been challenged, the physical soundness of the basic combination of the energy balance and aerodynamic approaches, has seldom been questioned. Reasonable modifications can be made more easily to the Penman formula than to the Thornthwaite formula owing to the more explicit physical nature of the former.

The third criterion is that daily values of AET are being sought. The examination of the daily data in section 4.3 indicated that the Penman formula was more sensitive to daily changes of weather. It is therefore more likely that the Penman method would give more accurate daily estimates of PET if it could be corrected for its overestimation.

On the basis of the discussion of all of these criteria it was decided to modify the Penman method.

4.4.f. <u>Modifications of the Penman Model</u>

Modifications were made to the Penman method as used in section 4.4.d. Both the radiation and the 'drying power' term were adjusted, and the combination of the two together was adapted to compensate for overestimation from October to March. An attempt was also made to adjust values of PET to those of AET where appropriate. The reasons for these specific alterations are discussed below where the modifications themselves are described. In addition to the known data (section 4.4.b.) computations of daily values of P and Bowen ratios were made. This was to ensure that the modifications gave rise to reasonable results with respect to the other heat balance components. The modifications adopted were as follows:-

1. Soil heat flow was added to the net radiation values over the whole study period. Tanner and Pelton (1960) showed that A should not be neglected except possibly during the growing season. The exception was made on the grounds that the A/Rn ratio was small at this time and that the ground would be completely covered by vegetation. In the Chilton Valley during the summer, the A/Rn ratio is relatively small, but there is no marked seasonal change in the area of vegetation cover comparable to that found where agricultural crops and/or deciduous vegetation are growing. The soil heat flow term was therefore incorporated with the net radiation value in all seasons.

2. A restriction on the total possible PET was applied on days when the 'drying power' term accounted for very much greater evaporation than the available energy term. Thornthwaite and Hare (1965) quote the Rn term as being typically six times greater than the drying power term. However, in the Chilton Valley föhn-like conditions often accompanied by low Rn values, low humidities and high winds, can lead to the 'drying power' term being the more important of the two. This situation was implied in section

4.3 for summer conditions, but is accentuated especially in winter when Rn values can be negative. There is no reason why the 'drying power' term should not have a larger value than the radiation term, but energy must be available for the change of state of water from liquid to vapour. To a certain extent this energy can be supplied by the downward flow of sensible heat which has been advected into the site. However, vertical temperature measurements suggest that the large downward flows of sensible heat were not available in the magnitude required by the original Penman model. As an example, on 20 July 1970 the original Penman model indicated values of LE that would require +166.9 ly day^{-1} of downward sensible heat flow. The average lapse rate between 3 m and 12 m for the day, based on 24, hourly readings, was +0.011°C m^{-1}. The average wind speed was 5.7 m sec^{-1}. An approximate estimate of P can be obtained from an empirical model of Frankenberger (1962) and this indicates that the maximum downward flow under these conditions could be only about +35 ly day^{-1}.

Owing to the possibility of advected heat, the 'drying power' term was not restricted directly. The condition was applied that if the total PET indicated by the Penman combination of radiation and 'drying power' terms, called for higher E than could be given by 1.20 of the available (Rn +A), then the value of PET was equal to the water equivalent of

1.20 (Rn + A). The value of 1.20 was selected after an examination of a review of literature, given by Chang (1968), on the fraction of Rn lost as LE. Moreover, the value allows for a certain amount of advected heat, and it gave rise to reasonable values of downward P in the winter with respect to lapse rate and wind conditions.

3. Penman's wind function was replaced by one suggested by Businger (1956). This is

$$f\ (u) = 1.2\ u\ \left[(1/k)\ \ln\ (z + z_o)/z_o\right]^{-2} \qquad 4.4.3.$$

This was applied in the manner described by Tanner and Pelton (1960) with an assumed roughness length of 9.0 cm. This roughness length is taken from values published in the latter reference, and is appropriate for a vegetation height of approximately 70 cm. It is also in agreement with measurements made near the recorder site (see section 5.4). Tanner and Pelton (1960) also showed that the Businger wind function was necessary in areas where advection of sensible heat was a common occurrence.

4. The assumptions concerning the saturation deficit term (Table 4.5) were adjusted so as to be more applicable to the whole year. Computations from data from other mountain stations in New Zealand (Coulter, 1967) showed that the average values of $(e_a - e_d)$ for the year, were 60% of the values in January and February. Since, in the preliminary Penman model assumed values of $(e_a - e_d)$ were based

on data collected in these months, new values (Table 4.5) of 0.6 times the original values, were adopted.

5. An attempt was made to obtain values of AET from those of PET under conditions of soil moisture deficit. The problem of determining when AET values are lower than PET values has been a matter of controversy for several decades. It has been reviewed by many authors (Penman, 1963, p.36; Thornthwaite and Hare, 1965; Chang, 1968). Veihmeyer and Hendrickson (1955) suggest that AET rates equal PET rates until all available soil water is gone. Thornthwaite and Mather (1954) propose that AET decreases as soil moisture tension increases. Other approaches (e.g. Penman, 1949) fall between these two points of view. Penman (1963 p.48) also points out that 'it may be ... neither of them is precise at any time, but the errors they carry with them are probably less important than other sources of uncertainty in their technical applications'. This is certainly relevant to the empirical approach to the problem that is followed in the present study. Two assumptions are made. Firstly it is assumed that the fraction of the estimated PET that is actually evapotranspired is proportional to the amounts of moisture available in the root zone of the soil. This, in fact, tends towards the Thornthwaite and Mather viewpoint. In a sense, such an approach is adopted by default, since data concerning the

actual wilting point of vegetation in the Chilton Valley are unavailable. The second assumption is that the amount of soil moisture available may be estimated by means of gypsum blocks. The limitations of such estimations are discussed in section 4.6.b.

On the basis of the above assumptions the following steps were taken. Examination of daily soil moisture values for summer (section 4.6.b., Fig. 4.12) indicated that under relatively high radiation conditions, PET rates could be sustained for about nine or ten days, but that if there was no rain after this time, soil moisture, and therefore presumably the AET/PET ratio, declined relatively quickly. Similar findings were reported by Black et.al. (1969a, 1969b) for work in Wisconsin. Based on the observed rate of decrease of soil moisture (Fig. 4.12) values of the AET/PET ratio (Table 4.6) were assumed. A rainfall value of 0.2 cm on any one day was selected as sufficient to allow a return to PET rates, although not necessarily to bring the soil back to field capacity. The values in Table 4.6 were applied, in the modified model, only during the months of relatively high Rn, October to March inclusive. Evidence from the computation of soil moisture storage (Table 4.10) using lysimeter data, indicated that sufficient soil moisture was available to permit PET in the remaining months of the study period.

6. A factor reducing estimates of PET by 0.1 was applied to the period October to March. Empirically, this was

TABLE 4.6

ASSUMED VALUES OF AET/PET RATIO FOR RAINLESS PERIODS BETWEEN OCTOBER AND MARCH IN THE STUDY YEAR

No. of Days After Rain of 0.2 cm on One Day	E act / E pot
1 - 8	1.00
9	0.95
10	0.925
11	0.9
12	0.8125
13	0.75
14	0.65
15	0.55
16	0.50
17	0.40
18	0.325
> 18	0.25

necessary since the use of lysimeter data indicated that the model, even with the above modifications, appeared to overestimate PET rates during this period. Also, for the period 23 January to 8 February, when the measured PET was only slightly more than 3.8 cm, the model gave a value of 4.5 cm. Overestimation by the method used here may be due to factors inherent in the Penman formula when applied to New Zealand, and/or to factors concerning the vegetation in the Chilton Valley.

Other studies, where the Penman formula has been applied to periods longer than three months, have shown the formula to overestimate. This is particularly true of the original Penman (1948) method which involved a two stage approach, where E from an open water surface was first calculated using a formula similar to equation 4.4.1., and then a factor was applied to obtain the PET value from the open water E value. Penman (1963, p.41) has noted a tendency for overestimation in this case. In New Zealand, Fitzgerald and Rickard (1960) using the Penman (1948) formula, found that a Penman factor which was 0.1 less than that applied in south east England, was more appropraite from September to January. Finkelstein (1961) also found it necessary to modify the Penman (1948) formula owing to its overestimation when applied in New Zealand.

The nature of the vegetation in the Chilton Valley should also be considered. Tussock grasses and the montane and sub-alpine plants are physiologically adapted to deal with dry conditions (Laing and Blackwell, 1964 p.41). Adaptations may take the form of the presence of relatively few stomata, which themselves may be protected from direct access to the atmosphere, and/or the partial or complete closing of stomata during daylight under certain conditions. Although such factors have not been specifically sutdied for the vegetation of the Chilton Valley, there is evidence that both may apply. A relatively low frequency and protection of the stomata of some of the flora in the Chilton Valley have been observed (Burrows, pers. comm., Zotov. pers. comm.). Like conditions are reported for similar plants elsewhere (Nebiker,1957; Metcalfe,1960 pp.135-145) Also, work on growth rates (Mark, 1965) and carbon dioxide exchange (Scott, 1970; Scott. et.al., 1970) of species found in the Chilton Valley indirectly suggests relatively low transpiration rates.

If vegetation factors, working to control ET, exist in the present situation, it is likely that they would be most effective in months when soil moisture deficit is likely. For this reason a reduction factor was applied to the present model during the months of likely soil moisture deficit (October to March). The value of the factor was selected in the light of the findings of Fitzgerald and Rickard (1960).

4.4.g. Results of the Modified Penman Model

The annual totals of PET and AET are 53.5 and 51.0 cm respectively, when the above modifications are applied to the original Penman model described in section 4.4.d. The annual total of AET estimated by the model may be compared with 56 cm measured by the lysimeters. With regard to the lysimeter measurements, random errors could account for 1.1 cm of the difference, and the effect of systematic errors in these measurements is not absolutely certain, except in the case of the fifth cause of error (see section 4.2 and Appendix G) which would lead to overestimation by the lysimeters. For these reasons, the model and lysimeter annual values are considered compatible.

The value of PET for the period January 23 to February 8 is 4.07 cm. Computation of the lysimeter soil moisture storage terms shows the maximum possible water loss by AET is exceeded on only two occasions (Table 4.10). Both apparent soil moisture deficits could be sustained if the field capacity of the field tanks had been computed for a 35 cm, instead of 30 cm, soil depth (see section 4.6.b.). Bowen ratio values (Table 5.1) are reasonable throughout the year. Comparison of the daily values given by the modified model, with those given when no assumptions were included (Table 4.3) shows that the sensitivity to daily weather changes is still maintained. A check on the effect of the modifications is obtained by computing, for the period January 20 to February 18, the relationship between Penman values of PET where no assumptions were used (PET_{NA}), and the values of AET given using the modifications (AET_M). The relationship is

$$AET_M = 0.56\ PET_{NA} + 0.38\ \text{mm day}^{-1} \qquad 4.4.4.$$

C.C. = 0.92 S.E.E. $= \pm 0.35$ mm day^{-1}

The results from the final modified model indicate AET rates equal PET rates for most of the period to which this relation applies (Fig. 4.6). Therefore the relatively high C.C. and low S.E.E. are indicative that the model assumptions, although lowering the value of PET given by equation 4.4.1., do not create large errors with regard to day to day variability.

4.4.h. Comments on the Model

The preceding sections have described the development and characteristics of the method used to obtain daily values of AET and consequently LE. The method has a number of limitations and the following are major areas to which further study should be directed.

1. The basic causes of overestimation of the Penman method should be sought. These may be associated with the way in which the method deals with advective influences, and the response of the vegetation in the Chilton Valley to the meteorological factors determining AET.
2. More daily saturation deficit data should be acquired and analysed, so that the assumptions concerning $(e_a - e_d)$ can be refined. Seasonal variation in the term should be included.
3. Further observations and refinement are required of the part of the model deriving AET values from PET values. The present assumption that the relation between AET and PET values is solely in terms of

soil moisture availability, and not related to other meteorological factors, may be incorrect (see Denmead and Shaw, 1962). Certainly, to use rainfall as the only index of soil moisture availability represents a gross simplification.

4. The applicability of different wind functions under conditions of non homogeneous vegetation cover should be studied.

5. The ratio of evaporation to transpiration is likely to change on a seasonal basis in the present location, and should be examined.

6. The assumption has been made that because the Penman method appears to be the method most sensitive to daily weather changes, it is more accurate in giving daily values of PET. In the absence of daily lysimeter, and/or accurate vapour flux measurements, the validity of this assumption cannot be properly tested. Although measurements of PET in other mid latitude sites (e.g. Davies and McCaughey, 1968; Högström, 1968) show the phenomenon to be sensitive to daily weather changes, this evidence is not entirely sufficient to conclude that the Penman method is more reliable than the Thornthwaite method in the present location. More accurate measurements should therefore be made to examine the relationship of PET rates to daily weather changes.

4.5 Values of Evapotranspiration and Evaporative Heat Loss

Daily values of AET and PET (Fig. 4.6) were computed using the modified Penman model described above. There was considerable day to day variation owing partly to the variability of Rn values, and partly to variations in both the value of the 'drying power' term and the soil moisture. The variability in PET and AET values, and consequently those of LE (Fig. 4.8), was most marked during the summer months when absolute values were at their maximum. There were several relatively short periods with consistently high PET values. In some of these periods, especially in November, AET rates were lower than PET rates. Monthly AET and LE values (Fig. 4.7 and Table 4.7) showed that LE varied in sympathy with AET as might have been expected.

The annual total of AET is 51 cm. The annual totals of PET for other relevant New Zealand stations (section 4.2) are compatible with this value. In addition, Finkelstein (1961) reports that the annual PET for a station at Harper river (within 45 km of the Chilton Valley) is 57 cm. This figure, preliminary data from Craigieburn (Morris ,1965), and the Chilton Valley records, all support Finkelstein's (1961) contention that 'the higher rainfall areas of Canterbury in and close to the Alps ... have evaporation about as high as the lower plains and the coast.'

The annual value of LE for the study period at the Chilton Valley was 30.1 kly. Comparison of monthly values of LE and Rn (Fig. 7.6) would suggest, at first sight, that there is a lag of LE values behind those of Rn. This situation might lead to the conclusion that LE was correlated more closely with mean air temperatures than with Rn values. However, in this case, relatively low values of LE in November and December are in fact due to AET values being lower than PET

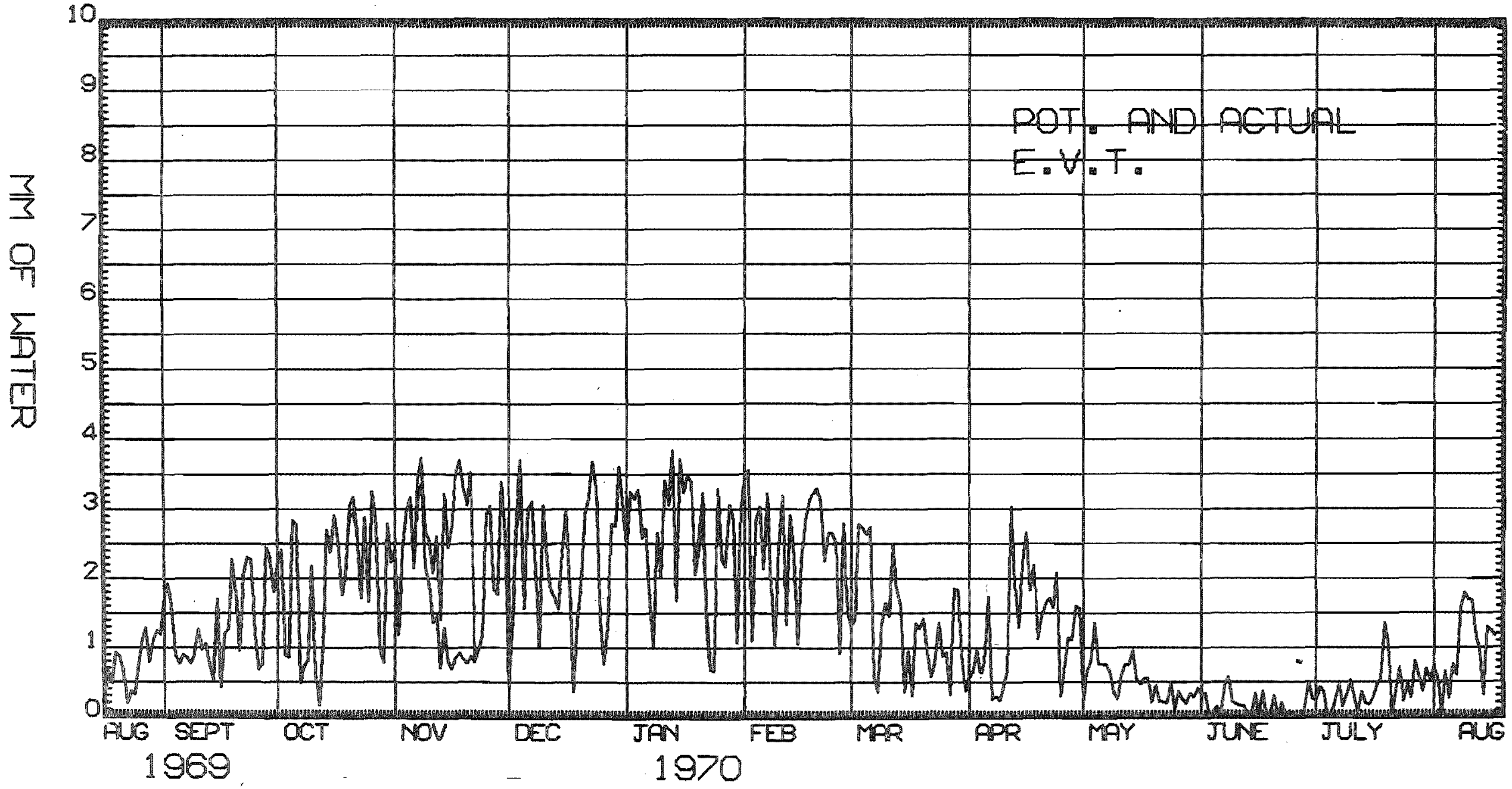


FIGURE 4.6: Daily values of PET and AET at the Chilton Valley for the study period, given by the modified Penman model. Values of AET as shown by the lower line on days when they do not equal values of PET

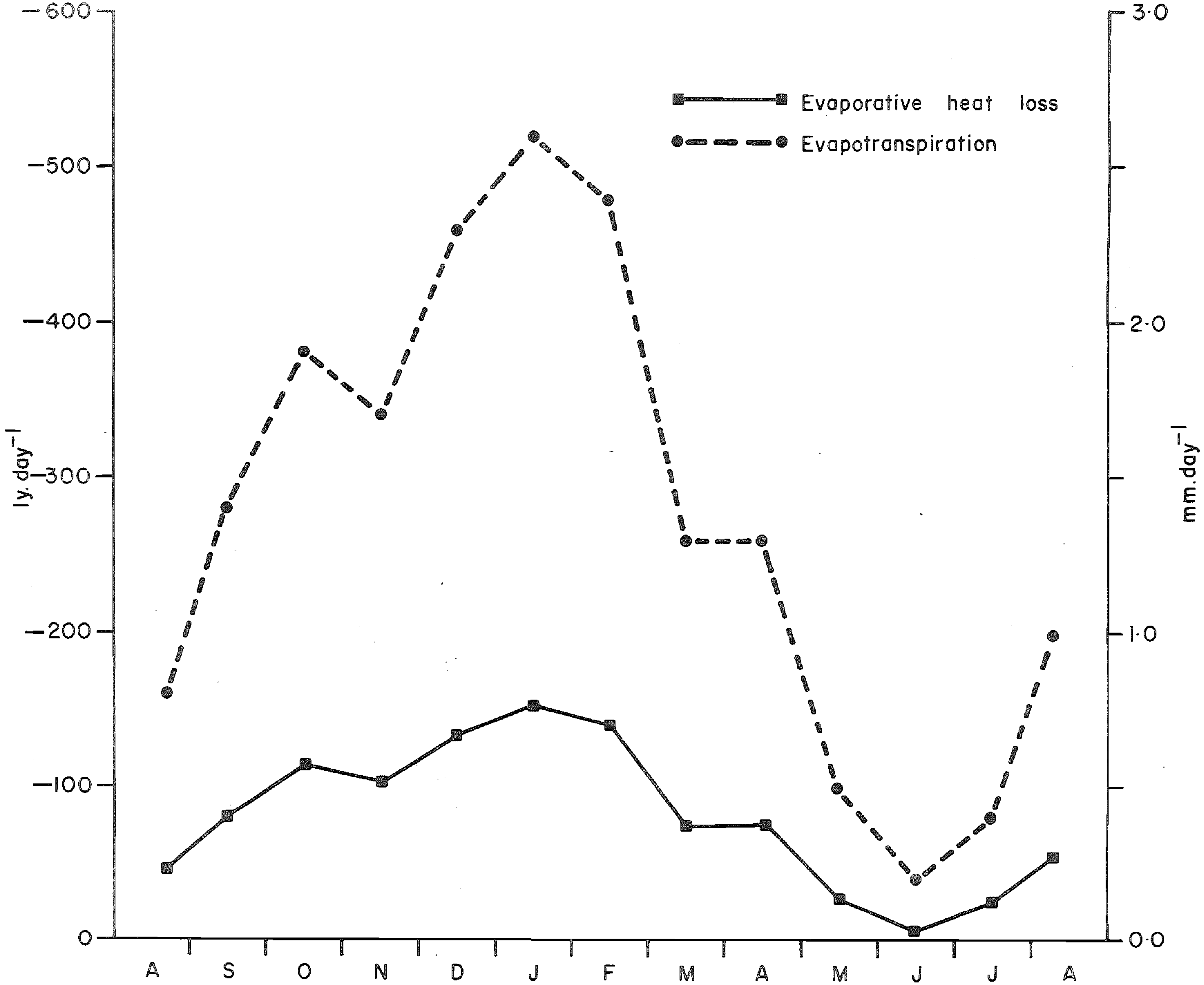


FIGURE 4.7: Monthly mean values of evaporative heat loss and actual evapotranspiration at the Chilton Valley during the study period

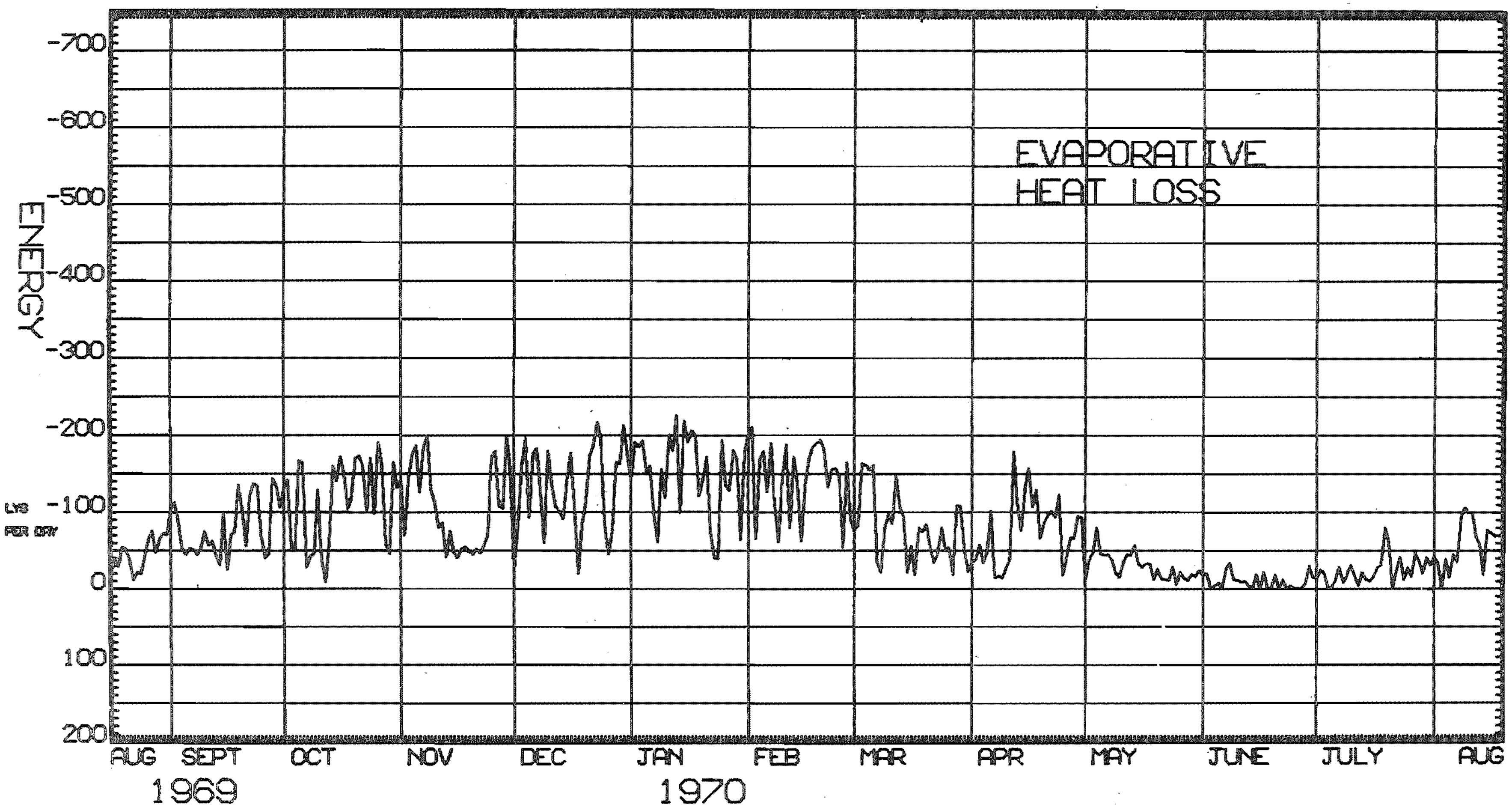


FIGURE 4.8: Daily values of evaporative heat loss at the Chilton Valley during the study period

TABLE 4.7

MONTHLY VALUES OF ACTUAL EVAPOTRANSPIRATION AND EVAPORATIVE HEAT LOSS

	Actual Evapotranspiration mm day^{-1}	Evaporative Heat Loss ly day^{-1}
August	0.8	- 48
September	1.4	- 80
October	1.9	-114
November	1.7	-102
December	2.3	-135
January	2.6	-151
February	2.4	-140
March	1.3	- 77
April	1.3	- 75
May	0.5	- 29
June	0.2	- 9
July	0.4	- 25
August	1.0	- 58

values during these months. At other times of the year LE rates appear to be closely associated with Rn values. Further discussion of AET and LE values is presented in Chapters 6 and 7.

4.6 The Wetting and Drying of the Soil

4.6.a. Introduction

The thermal response of the soil to the input of rainwater was considered in section 3.4. Equally important to biological and geomorphological processes, is the manner and frequency of soil wetting and drying. Frequency of soil wetting at the Chilton Valley has been discussed elsewhere (Greenland and Owens, 1967). During the period January 20 to February 26, data were collected which show how the value of soil moisture responds to wetting and drying over a short period. These results will be presented and then followed by an examination of soil moisture on a long term scale.

4.6.b. Short Period Investigations

Four gypsum blocks were used to measure values of soil moisture. The blocks, which are described elsewhere (Bouyoucos and Mick, 1940; Bouyoucos, 1952), were for agricultural use and values of soil moisture were given in terms of 'percentage available moisture'. The blocks were calibrated among themselves (Table 4.8) before being placed in and near the lysimeters. A consistent rank order was found among the blocks, the order being, from highest to lowest block C, D, A and B, respectively. The maximum difference in values given by the blocks was 6.0%. It was concluded that the blocks were consistent in relation to each other but not

TABLE 4.8

CALIBRATION OF THE SOIL MOISTURE BLOCKS AMONG THEMSELVES. UNITS ARE 'PERCENTAGE AVAILABLE MOISTURE'

Date	Block A	Block B	Block C	Block D	Maximum Difference
4.9.69	73.0	71.0	75.0	73.5	4.0
5.9.69	74.0	73.0	77.0	75.5	4.0
23.9.69	80.0	78.0	84.0	81.0	6.0
Averages	75.7	74.0	78.7	76.7	
	74.8		77.7		

TABLE 4.9

PERCOLATION DATA FROM MODIFIED RUNOFF PLOT

Date	Rain cms	Percolation cms	Time from Start of rain to start of Percolation hrs	Time from cessation of rain to cessation of percolation hrs
1970				
21-23 Jan.	6.70	0.26	6.9	27.4
29 Jan.	0.47	None	-	-
4 Feb.	0.52	None	-	-
7-8 Feb.	1.75	Trace	6.1	-1.0
13 Feb.	0.20	None	-	-

similarly sensitive.

It was hoped to calibrate the block recordings against gravimetric measurements of soil moisture. However, the resultant scatter of points (Fig. 4.9) was too great to give any confidence in the relationship. The correlation coefficient of the two sets of measurements was 0.64. The reasons for the poor relationship include the difficulty of taking soil samples near the blocks without disturbing them, the fact that the samples could not be analysed immediately, the low range of moisture contents, and the temperature dependence of the blocks themselves. Comparison of the results from the blocks with soil moisture storage estimated by the Thornthwaite method (Fig. 4.10) showed less scatter. In making this comparison it was apparent that the Thornthwaite soil moisture estimates gave a linear decline during drying periods, but that the blocks showed soil moisture to be constant for most of the time except at the end of drying periods when a large drop occurred. Also, when the Thornthwaite estimates show field capacity has been reached, there is some lag before the block readings rise up to their maximum.

Uncertainty in the measurements with the gypsum blocks therefore exists. The blocks may be used only for giving relative values, but such values are consistent and useful for pointing out trends in soil moisture. The blocks were placed at 10 cm and 20 cm beneath the surface, inside and just outside the field tank of lysimeter 2.

Wetting periods are examined first, and one of these occurred on 21-23 January 1970 (Fig. 4.11), when 7.5 cm of rain was recorded by the lysimeter gauges. The rain had an effect relatively quickly at the 10 cm levels, but there was a delay

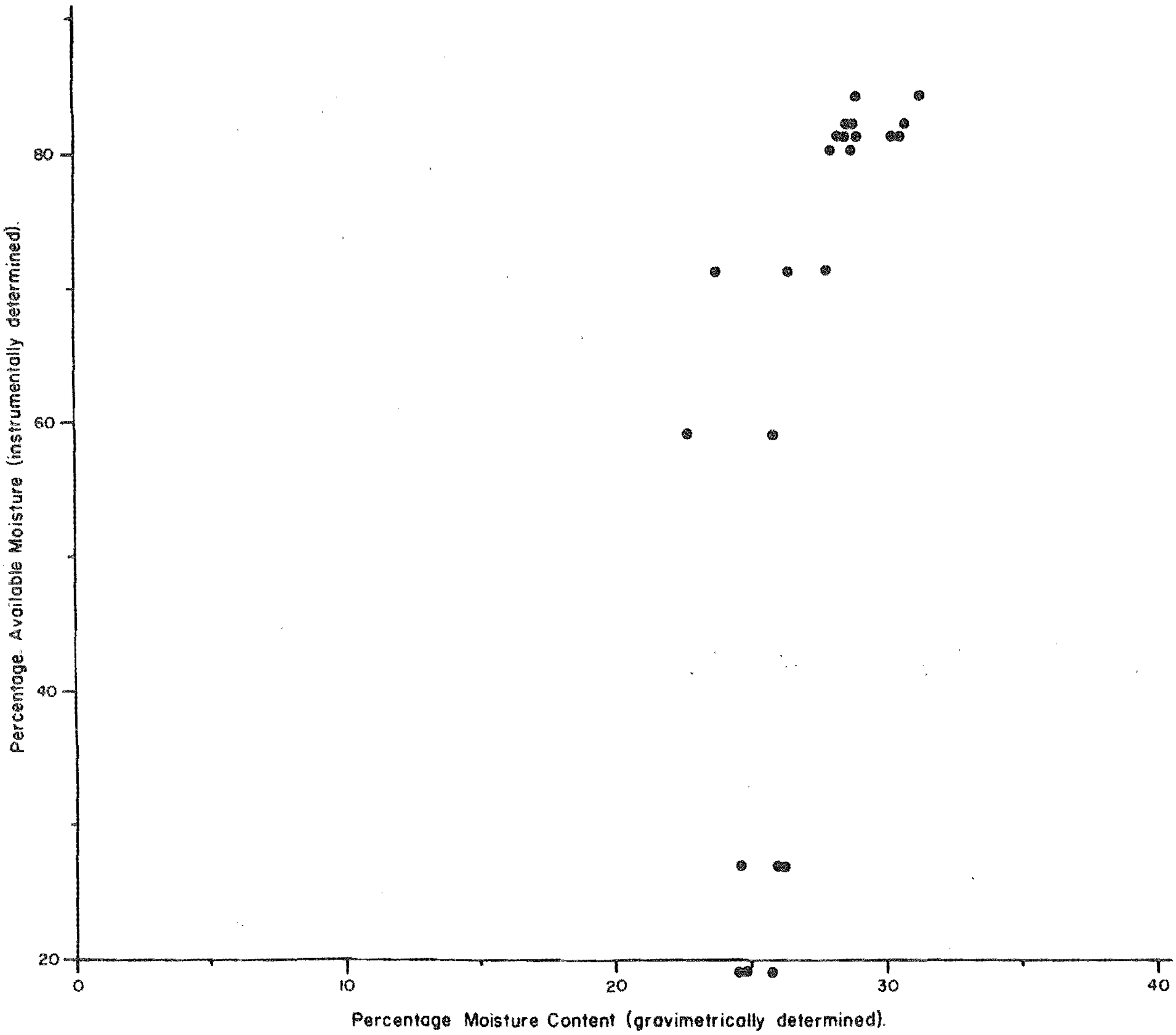


FIGURE 4.9: Values of percentage available soil moisture given by gypsum blocks and percentage moisture content determined gravimetrically

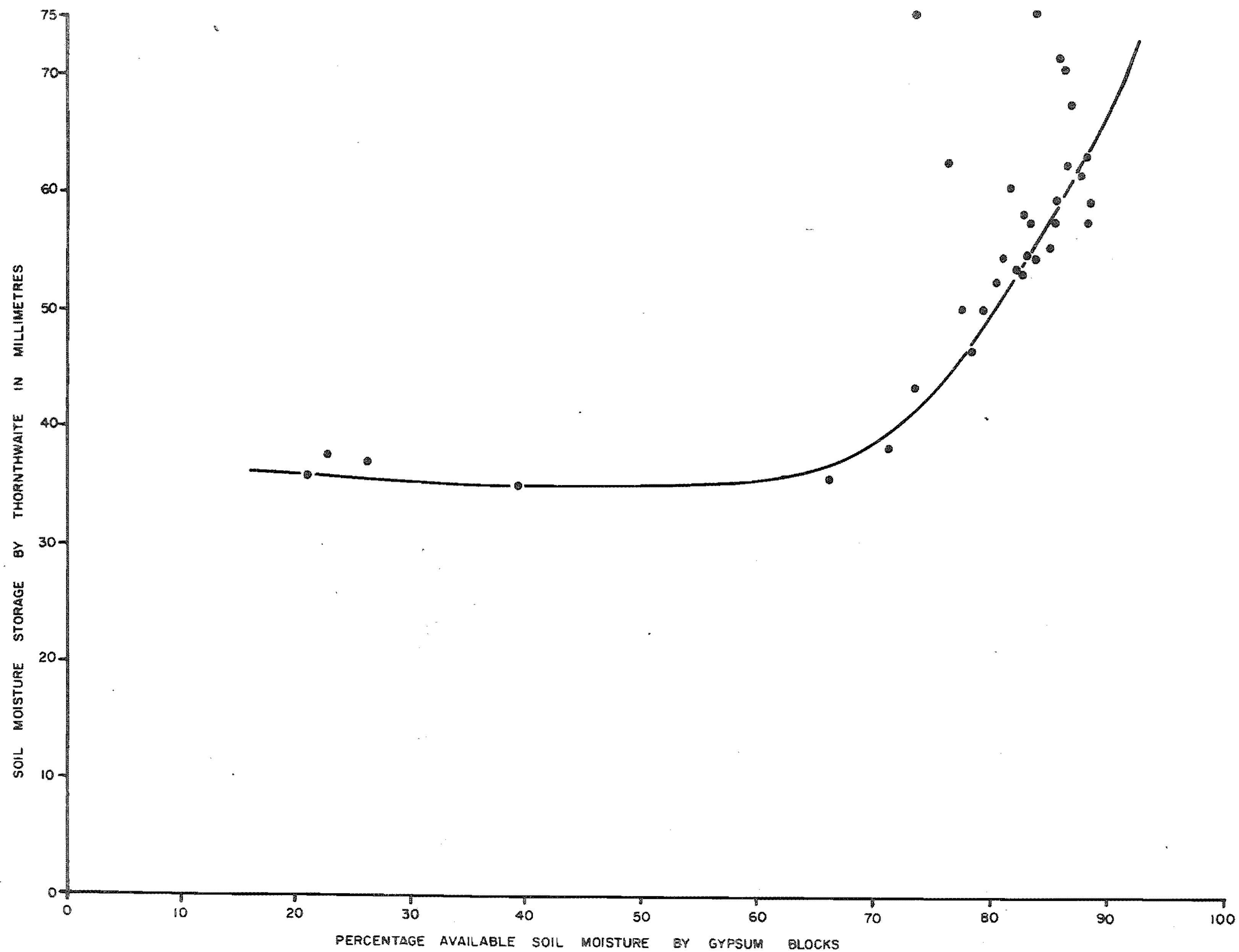
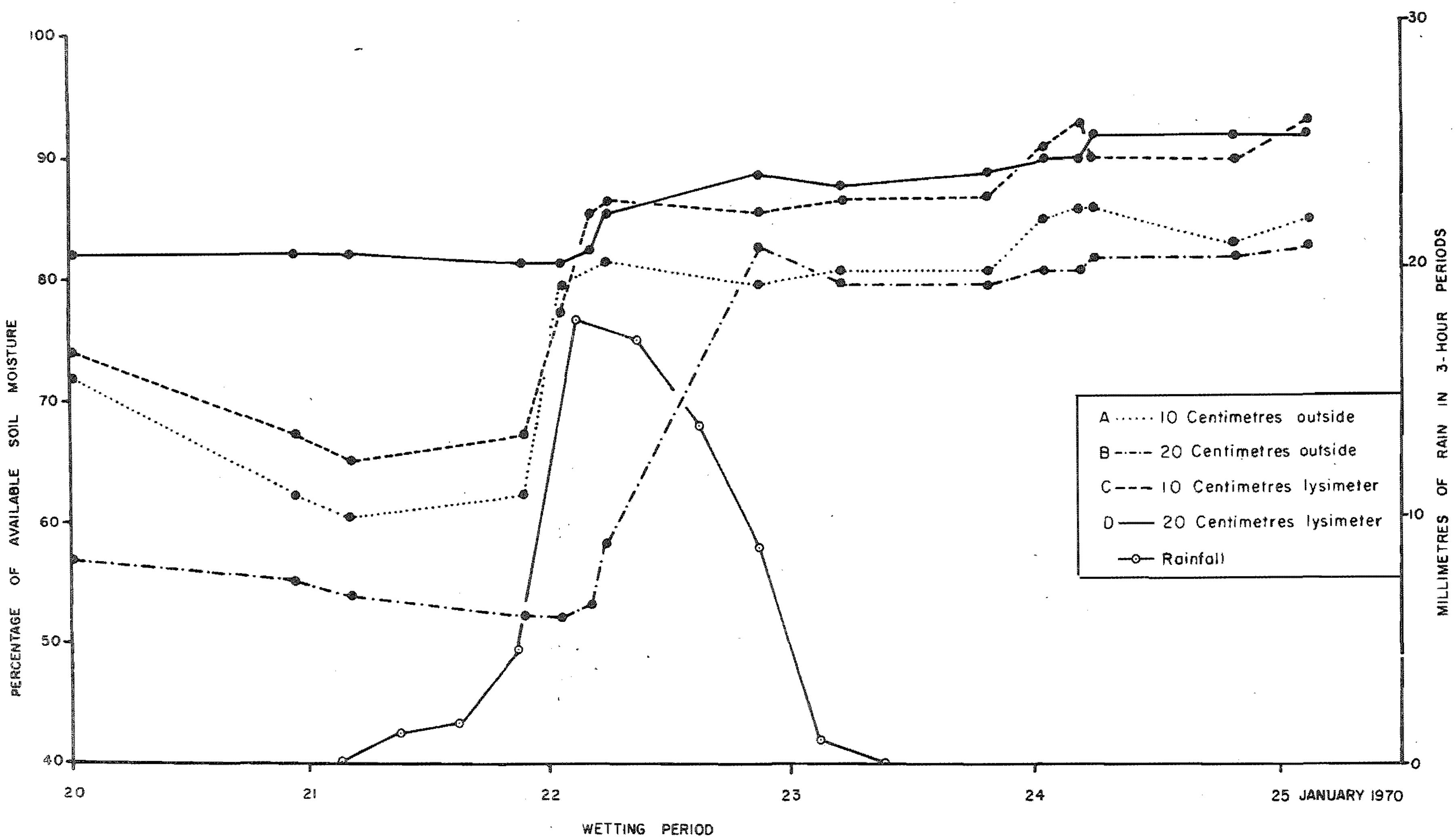


FIGURE 4.10: Values of percentage available soil moisture given by gypsum blocks and soil moisture storage given by the Thornthwaite method. (Thornthwaite and Mather, 1957)

FIGURE 4.11: Values of percentage available soil moisture and rainfall during a wetting period (20-25 January 1970)



of about 13 hrs before the effect was noted outside the field tank. The period starting at 1100 hr on the 22 January, had the highest intensity rainfall (3 mm hr^{-1}). This resulted in an almost immediate response at the 10 cm level, presumably partly due to previous wetting, but at the 20 cm level even inside the field tank, a steep rise in the value of the available soil moisture occurred after about 7 hr. The high value of soil moisture at 20 cm inside the field tank before the commencement of the rain is further evidence that the field tank retained water that would otherwise have infiltrated to lower levels.

Additional data on the rate of soil wetting may also be obtained from the modified runoff plot described in section 1.5. In the particular storm examined above, percolation started from a depth of 60 cm, 6.9 hr after the commencement of the rain (Table 4.9). When the rain stopped percolation continued for another 27.4 hr. Later, lighter rainfalls caused no percolation, but one of 1.75 cm caused a slight amount of percolation which actually ended before the rain stopped.

Although more data are needed in order to give greater confidence, two tentative conclusions may be drawn from the above evidence. Firstly, a lag effect is present in the wetting of the lower layers of the soil. This is analogous to that in the heating of the lower levels described earlier (section 3.4). Secondly, despite the lag effect, following heavy rainfall conditions, such as those described above, some moisture can percolate to a depth of 60 cm within 7 hours after the onset of the rain, and much of the water that will percolate to below 60 cm does so within 24 hrs of the cessation of the rain. Following heavy rainfalls therefore, much of the water will be removed from the root zone in a short space of time.

The drying period that followed the rainfall of 21-23 January 1970 was also examined (Fig. 4.12). During the first part of the drying period a decline in soil moisture occurred only after nine or ten days. Light rainfalls, such as that on February 3, helped to keep soil moisture high at the 10 cm level but did not arrest drying at the 20 cm level, particularly outside the field tank. Over the whole period between January 24 and February 26, the soil moisture values, as shown by the blocks, present an interesting picture of the drying of the soil. Despite high Rn and AET values during the period, with the help of light rainfalls, soil moisture values remain high for all except the last part of the drying period. On the last few days the soil moisture is depleted quickly and is not sustained by the light rain on February 21. Therefore, despite the relatively rapid percolation rates described above, it appears that there can be available moisture within the root zone for a large part of a drying period. This will probably influence the value of the AET/PET ratio.

The field capacity of lysimeter 2 was investigated on 28 and 29 January 1971. Prior to this period the gypsum blocks indicated zero soil moisture at both levels. Water was added to the surface of the field tank until percolation flowed freely. The tank was left for a day before more water was added. This percolated almost immediately. Comparison of the input and output of water, after it had been allowed to

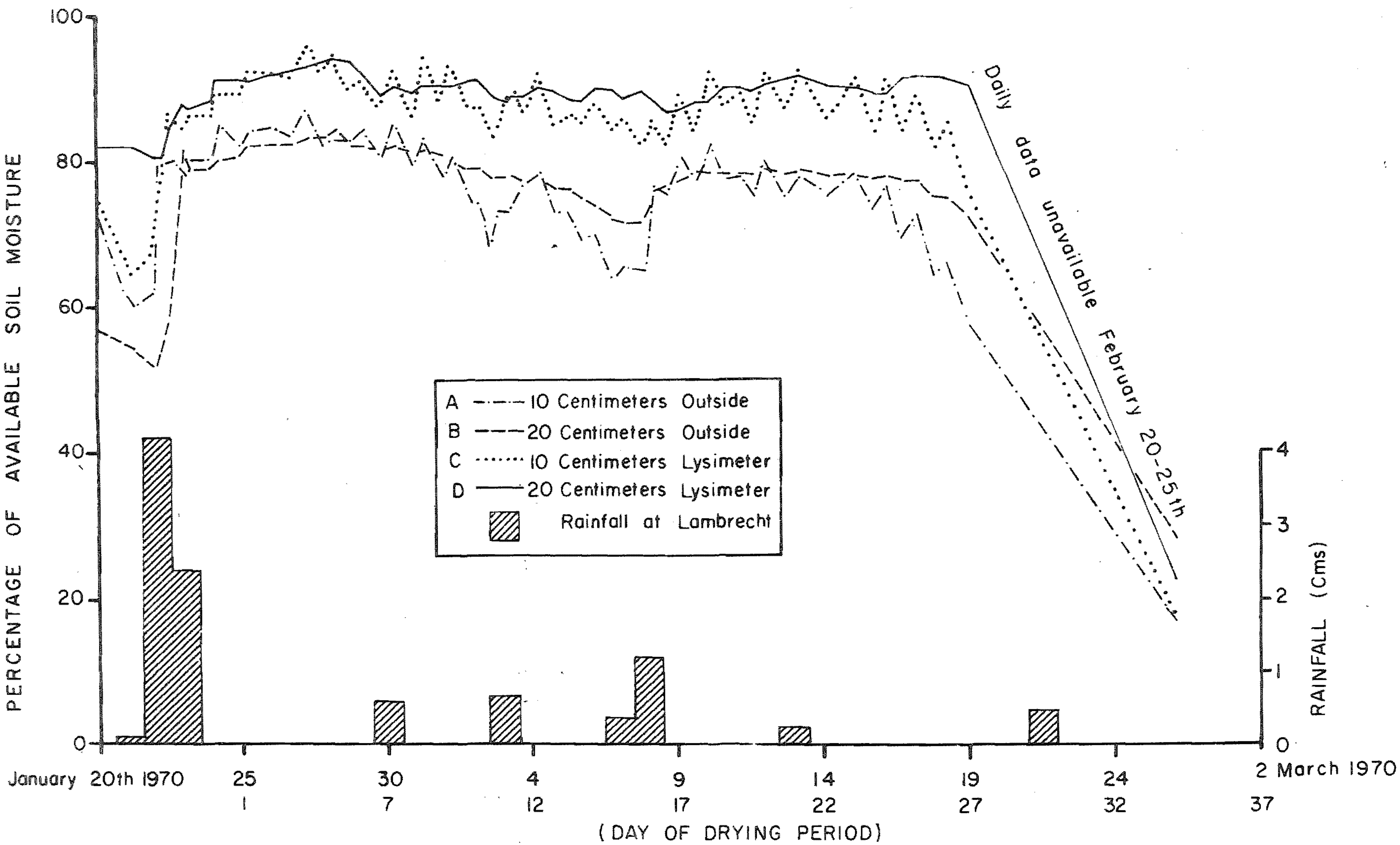


FIGURE 4.12: Values of percentage available soil moisture and rainfall during a drying period (20 January - 26 February 1970)

drain, and consideration of the volume of soil contained in the field tank, showed that the water retention capacity of the tank was 0.175 ± 0.025 gm for each cubic cm of soil plus air contained in the tank. This compares favourably with values obtained by Archer and Collett (1970) of 0.13 to 0.21 gm cc^{-1} for similar high country soils. When the tank is at its minimum soil moisture content therefore, it would take a rainfall of between 5.25 ± 0.75 cm, for a 30 cm soil depth, and 6.10 ± 0.75 cm, for a 35 cm soil depth, to return the tank to field capacity. In the computation of soil moisture storage using the lysimeter data (Table 4.10), the lower values of field capacity and soil depth were used since the lower 5 cm of the field tanks were filled with gravel as explained earlier (section 4.2). In the computation of the daily water balance by the Thornthwaite method (sections 4.4.c. and 4.6.c.) a water retention value of 0.15 gm cc^{-1} was used since it was the nearest to the Chilton Valley conditions that is tabulated by Thornthwaite and Mather (1957).

4.6.c. Long Period Investigations

Two approaches were used to investigate the values of soil moisture throughout the year. First, the lysimeter data were used with the modified Penman model values of AET to give the soil moisture content at the end of the periods between lysimeter readings. In this study, the soil moisture value was obtained by using sequential solutions of the water balance equation of the field tank. Secondly, soil moisture storage was computed by the Thornthwaite book-keeping method (Thornthwaite and Mather, 1957). The latter may be used where only temperature and rainfall data are available. Because of its wide applicability therefore, it is of interest to compare soil moisture values derived from the Thornthwaite technique with those based on the lysimeter and modified Penman data.

Lysimeter data were used in the following way to give soil moisture storage values on the dates when lysimeter measurements were taken. Consider the water balance equation of the field tank at the end of the first period of measurement ($i = 1$). At the beginning of the period there is a soil moisture storage S_o. During the period additional water is given by the incoming rainfall minus the outgoing percolation ($Ra_1 - Pe_1$). Subtracted from S_o is the AET value for the period, AET_1. At the end of the period a new value of soil moisture storage is given by

$$S_1 = S_o + (Ra_1 - Pe_1) - AET_1 \qquad \text{—— 4.6.1.}$$

S_1 now becomes the initial soil moisture storage for the second period ($i = 2$) and the storage at the end of this period is similarly given by

$$S_2 = S_1 + (Ra_2 - Pe_2) - AET_2 \qquad \text{—— 4.6.2.}$$

and so on. In order to solve this series of equations for the study year, use was made of the following points. S_o was taken to be the field capacity value (5.3 cm). This is reasonable for the time of the year (August 15) and considering the previous rainfall values (6.2 cm in the preceding 14 days). Ra_i and Pe_i were measured for each period, and the values of AET_i were estimated by the methods described earlier (sections 4.4.c., 4.4.d., and 4.4.f.). When the lysimeters were read just following, or during, a rain period a value of S_i higher than that of the field capacity could be obtained. In such cases the value of gravity water was neglected for the calculation of S_i which became 5.3 cm, but in the field situation the extra water entered into the soil and subsequently into the calculation of S for the following period, S_{i+1}.

Soil moisture storage (Table 4.10) was computed by means of the above procedure. This water balance approach was also used in testing (1) AET values given by Thornthwaite method discussed in section 4.4.c. (see cols. 2 and 5), (2) PET values given by the first Penman model described in section 4.4.d. (see cols 3 and 6), and (3) AET values given by the modified Penman model of section 4.4.f. (see cols 4 and 7). Soil moisture values resulting from the latter method are discussed below.

Field capacity is seen to have been attained at the beginning and the end of the study year. With the occurrence of precipitation throughout the year, most of the periods had at least half the field capacity of the field tanks. There were two notable exceptions to this. One was the period in early November when 23 consecutive rainless days which had high values of Rn, exhausted the soil moisture. Despite rain in early December low soil moisture storage persisted until the precipitation of mid and late December (Fig. 4.13). (An explanation of the slightly negative soil moisture values of 18 November and 15 December is given in section 4.4.g.). A second period when low soil moisture was apparent is indicated by the value at February 25. This came at the end

TABLE 4.10

COMPUTATION OF SOIL MOISTURE STORAGE AT END OF PERIODS WHEN LYSIMETER READINGS WERE TAKEN. VALUES IN cm

		EVAPOTRANSPIRATION			LYSIMETER SOIL STORAGE		
Column No.	1	2	3	4	5	6	7
Date at end of Period	Rainfall-Percolation	Thornthwaite A E T	1st Penman Model PET	Modified Penman Model AET	Thornthwaite	1st Penman Model	Modified Penman Model
10.9.69	4.4	2.1	3.7	2.5	5.3	5.3	5.3
24.9.69.	1.3	1.3	2.7	2.0	5.3	3.9	4.6
3.10.69	0.1	1.1	2.2	1.4	4.3	1.8	3.3
22.10.69	2.6	2.4	6.7	3.6	4.5	-2.3	2.3
4.11.69	1.9	1.8	5.0	2.9	4.6	-5.4	1.3
18.11.69	0.0	3.6	6.6	2.2	1.0	-12.0	-0.9
2.12.69	3.7	3.7	4.4	2.5	1.0	-12.7	0.3
15.12.69	2.1	3.6	4.7	3.0	-0.5	-15.3	-0.6
27.12.69	7.0	3.6	3.4	2.5	2.9	-11.7	3.9
11.1.70	2.8	5.3	5.9	4.1	0.4	- 14.8	2.6
19.1.70	2.3	2.6	3.8	2.4	0.1	-16.3	2.5
30.1.70	3.9	3.8	3.8	2.3	0.2	-16.2	4.1
15.2.70	3.3	3.5	6.0	3.8	0.0	-18.9	3.6
25.2.70	0.5	2.8	4.0	2.7	-2.3	-22.4	1.4
10.3.70	4.5	3.4	3.5	2.3	-1.2	-21.4	3.6
24.3.70	2.2	2.5	2.9	1.6	-1.5	-22.1	4.2
8.4.70	3.0	2.3	2.3	1.2	-0.8	-21.4	5.3
22.4.70	0.0	1.9	4.7	2.4	-2.7	-26.1	3.6
7.5.70	0.9	1.4	2.9	1.4	-3.2	-28.1	3.1
3.6.70	3.7	1.6	2.7	1.0	-1.1	-27.1	5.3
16.6.70	1.3	0.6	0.8	0.3	-0.4	-26.6	5.3
30.6.70	0.0	0.8	1.7	0.2	-1.2	-28.3	5.1
16.8.70	4.2	4.0	9.8	2.9	-1.0	-33.9	5.3

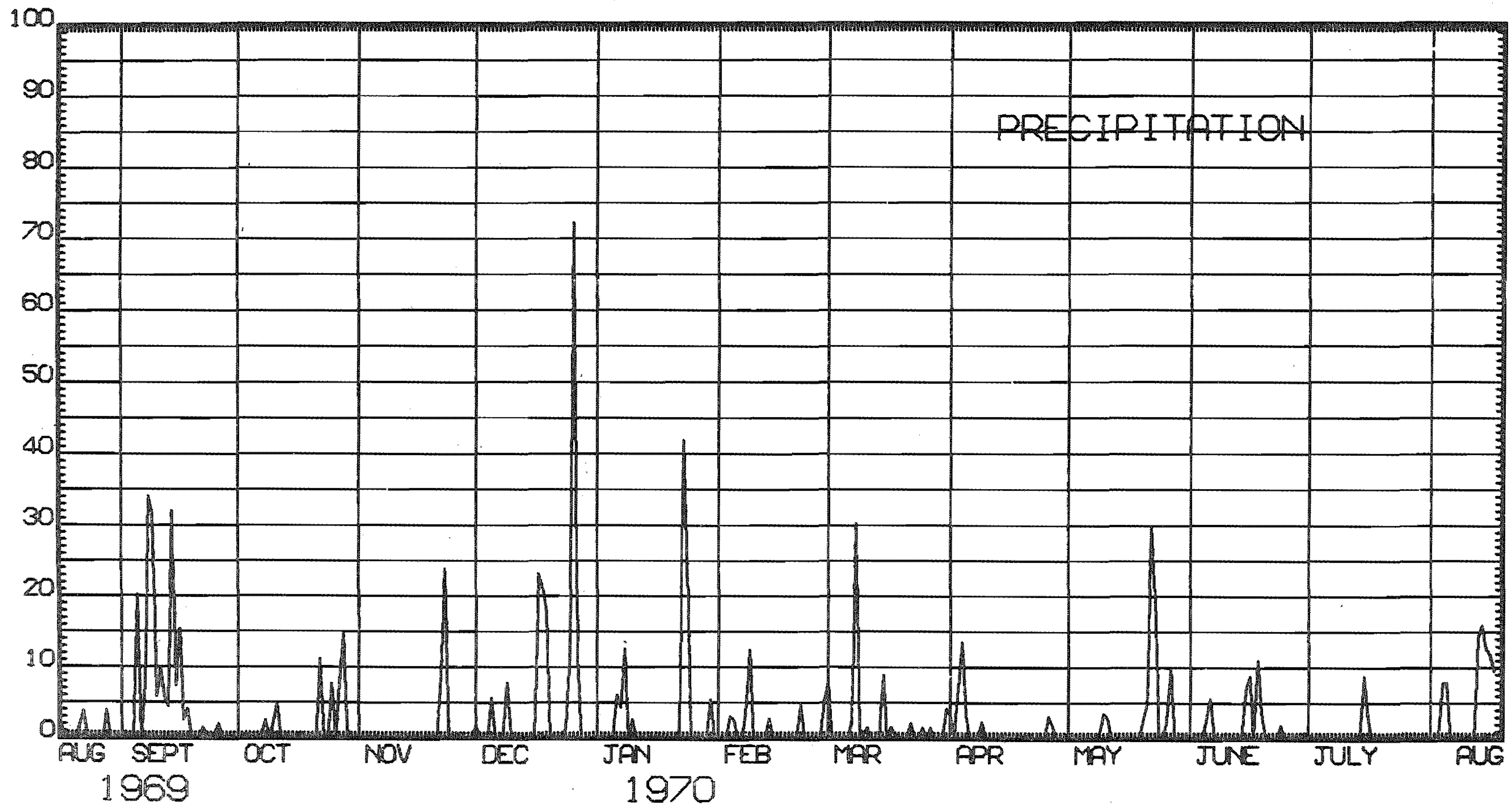


FIGURE 4.13: Daily values of precipitation recorded by the Lambrecht raingauge at the Chilton Valley during the study period

of the drying period described in the previous section and was relieved by rain on March 7.

Soil moisture storage was also computed by means of the Thornthwaite book-keeping method (Fig. 4.14). For a reason given earlier (section 4.4.c.) values of possible water retention of 75 mm and a root depth of 50 cm were selected for this computation. The most obvious feature illustrated by the results of the book-keeping method is the effect of the heavy summer rainfalls which restored soil moisture to field capacity three times during December and January. Low soil moisture values of mid January, March and early May are also noteworthy. The book-keeping method, in agreement with the modified Penman model-water balance method, shows field capacity values to have occurred at the beginning and at the end of the study year. An analysis of soil moisture storage given by the book-keeping method (Table 4.11) shows that over half of the days of the study year had soil moisture values greater than 80% of the field capacity value, and 93% of the days had more than half of the value selected for maximum soil water retention.

Comparison between the modified Penman-water balance soil moisture values (Table 4.10, col 7) and the corresponding values from the book-keeping method (Fig. 4.14) shows that the two methods indicate similar trends. The main difference is in the value used for field capacity. It therefore appears that the Thornthwaite method would be satisfactory for monitoring soil moisture in this location, if it was possible to use an accurate value of maximum water retention.

Both methods indicate that values of soil moisture

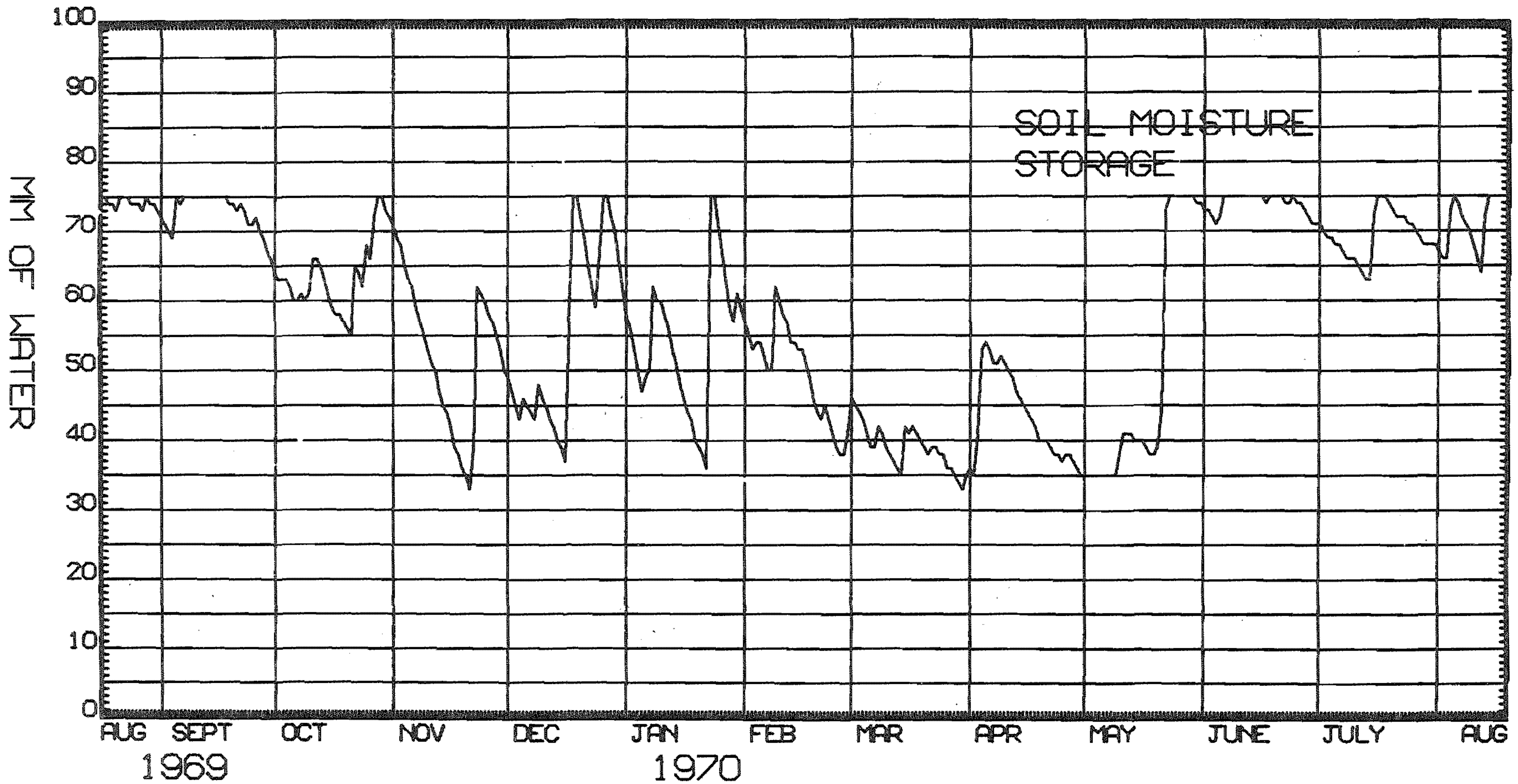


FIGURE 4.14: Daily values of soil moisture storage at the Chilton Valley during the study year, given by the Thornthwaite method (Thornthwaite and Mather, 1957)

TABLE 4.11

ANALYSIS OF DAILY SOIL MOISTURE VALUES GIVEN BY THE THORNTHWAITE BOOK-KEEPING METHOD

1 Percentage of Field Capacity	2 Soil Moisture Equivalent in mm Water	3 No.of Days in Each Category	4 Accumulative No. of Days With Soil Moisture ⩾ Values in Cols. 1 & 2	5 Percentage of Total that Nos. of Days in Col. 4 Represent
⩾ 0	0	0	365	100
⩾ 10	7.5	0	365	100
⩾ 20	15.0	0	365	100
⩾ 30	22.5	0	365	100
⩾ 40	30.0	27	365	100
⩾ 50	37.5	69	338	93
⩾ 60	45.0	39	269	74
⩾ 70	52.5	38	230	63
⩾ 80	60.0	55	192	53
⩾ 90	67.5	80	137	38
100	75.0	57	57	16

exceeded half the value of field capacity for the majority of the study year, although there were some notable exceptions in months of high Rn values. If it is assumed that a high soil moisture status is related to rates of transpiration, then it is probable that AET rates equalled PET rates for a large part of the year.

4.7 Summary

The main conclusions and observations that arise from this study of evapotranspiration and evaporative heat loss are as follows:-

1. Measurements of annual AET using non-weighing lysimeters showed a value of 56 cm during the study year at the Chilton Valley site. This compared favourably with estimates of evaporative water loss made in other areas of the High Country. Examination of evaporation data for the Chilton Valley and other High Country stations suggests that soil moisture deficits may occur in summer at lower altitude stations.

2. An examination of PET estimates for a short period in summer showed that the Thornthwaite method of estimating PET is satisfactory for short period totals, but is not entirely adequate for giving daily values. On the other hand, the Penman formula tends to overestimate PET values but appears more sensitive to daily weather changes.

3. After examining the Thornthwaite and Penman approaches for estimating daily AET and PET values respectively, it was decided to modify the latter approach in order to

obtain daily values of AET for the Chilton Valley. In the course of the examination of the applicability of the Penman formula, a tendency for it to overestimate was apparent. It has been suggested that the over-estimation may be caused by the resistance of some of the flora at the location to transpiration. The final modified Penman model indicated an annual value of AET of 51 cm compared with the value of 56 cm from the lysimeters.

4. The value of AET from the modified model is equivalent to 30.1 kly of LE. On a monthly basis the value of LE showed similar trends to that of Rn, except at times of large soil moisture deficit, which occurred particularly in November and December of the study year.

5. A study of soil wetting and drying was made. During wetting periods, despite a lag in the effect of increased soil moisture being felt at lower levels, there was evidence that much of the incoming rain water, during heavy falls, was removed from the root zone within 24 hr of the cessation of the rain. During a drying period it was found that, with the help of light rainfalls, soil moisture values at 10 cm and 20 cm were maintained for 9 or 10 days under conditions of high radiant input. However, soil moisture values declined steeply at the end of the period. The field capacity of the lysimeters was found to be between 5.25 ± 0.75 cm and 6.10 ± 0.75 cm. Calculation of soil moisture storage by the Thornthwaite book-keeping method showed that soil moisture remained at above half of

the selected maximum water retention value for 93% of the days of the study year.

CHAPTER FIVE

SENSIBLE HEAT FLOW

5.1 Introduction

In this chapter computed values of sensible heat flow are presented. Studies of the related phenomena of wind velocities, profiles and fields, and the frequency of advection of heat are also reported.

The direct measurement of sensible heat flow on the time scales considered in this study is prohibited by two principal factors. Firstly, there are the practical difficulties of insufficient data and the quality of the available data. A practical equation in the aerodynamic method of determining the vertical flux of heat takes the form

$$P = -\rho c_p k^2 \left(\frac{\Delta u \, \Delta T}{\ell n \, z_2 / z_1} \right) \qquad \text{—— 5.1.1.}$$

In the application of this formula, although data on ρ, c_p, k, and ΔT are available, the vertical variation of wind velocity Δu is not.

A second factor is that even if high quality data were available, most aerodynamic theory applies only to a large, flat, homogeneous surface and not to the kind of terrain and surface in the Chilton Valley. There have been several initial studies on the effect of changing roughness length on turbulent transfers and wind relationships (e.g. Rider et.al. 1963; Blackadar et.al. 1967), and Blom and Wartena (1969) have developed a method of dealing with more than three changes of roughness length. However, many studies (e.g. Taylor, 1969) show that even under neutral stability, a very long fetch is required for a return to true equilibrium flow conditions following a change of surface roughness. Therefore the application of even elementary aerodynamic principles must be approached with caution in the present location. A preliminary examination is made (in section 5.4) with regard to the applicability of some aspects of aerodynamical theory to the Chilton Valley site.

The factors mentioned above force an estimation of P to be made as the residual of the other energy balance components. Budyko (1958 p.60) points out that such a procedure has been used in many investigations and is satisfactory as long as P is relatively large compared with the principal terms of the heat balance, and in particular Rn. Results in the present study (Table 6.1, Appendix F) indicate this provision to be met at almost all times.

Although there are difficulties in measuring the values of P directly, phenomena related to P may be examined. Wind velocities are an integral part of equation 5.1.1., and give an indication of the relative dominance of free and forced sensible heat transfer. Wind directions and velocities are related in this location to the horizontal advection of air of particular thermal characteristics. Air temperatures may be used not only to indicate atmospheric stability, but also the frequency of advection at the site. Aspects of both wind and temperature variations are examined in sections 5.3 - 5.5.

5.2 Computed Sensible Heat Flow

The most outstanding feature of monthly mean values of sensible heat transfer for the study period (Table 5.1 and Fig. 5.1) is the high value for November when water shortage decreased evaporative heat loss. This value disrupts an overall seasonal trend that is common to many mid latitude climates (see for example Fig. 7.7). Summer values of P were approximately two thirds of the values of LE. The winter values of P fell off markedly to a minimum of only -1 ly day^{-1} in June. On many days during June there were positive flows toward the surface. Other stations, at even lower latitudes (e.g. Lisbon, Portugal, latitude 39°N, Fig. 7.7) show positive values of P in winter, and it is probable that monthly mean values of P at the Chilton Valley are also directed towards the surface during this season in some years. The annual mean value of P for the study year is -58 ly day^{-1}. This is identical to that of Astoria, Oregon at latitude 46°N (see Table 7.8). More detailed comparisons with other stations are made in section 7.4.

Monthly mean values of the Bowen ratio, P/LE, are also shown in Table 5.1. Values ranged from 0.07 for the first part of August 1970 to 1.50 for November 1969. The latter again reflects the relatively low LE value at that time. Bowen ratios in other mid-latitude climates are typically 0.1 over moist soils but can be larger when the soil moisture is below field capacity (Thornthwaite and Hare, 1965). Davies and McCaughey (1968) report summer values of 0.1 - 0.4 for

TABLE 5.1

MONTHLY MEAN VALUES OF SENSIBLE HEAT TRANSFER AND BOWEN RATIO

	Sensible Heat Transfer ly day^{-1}	Bowen Ratio P/LE
August	- 46	0.96
September	- 54	0.68
October	- 82	0.72
November	-153	1.50
December	-105	0.78
January	-101	0.67
February	- 87	0.62
March	- 56	0.73
April	- 12	0.16
May	- 22	0.76
June	- 1	0.11
July	- 3	0.12
August	- 4	0.07

TABLE 5.2

MEAN DAILY WIND VALUES IN m sec^{-1}

August	2.6	March	0.8
September	3.9	April	2.5
October	4.3	May	1.1
November	2.5	June	1.8
December	1.5	July	2.8
January	2.3	August	3.5
February	1.7	Year	2.4

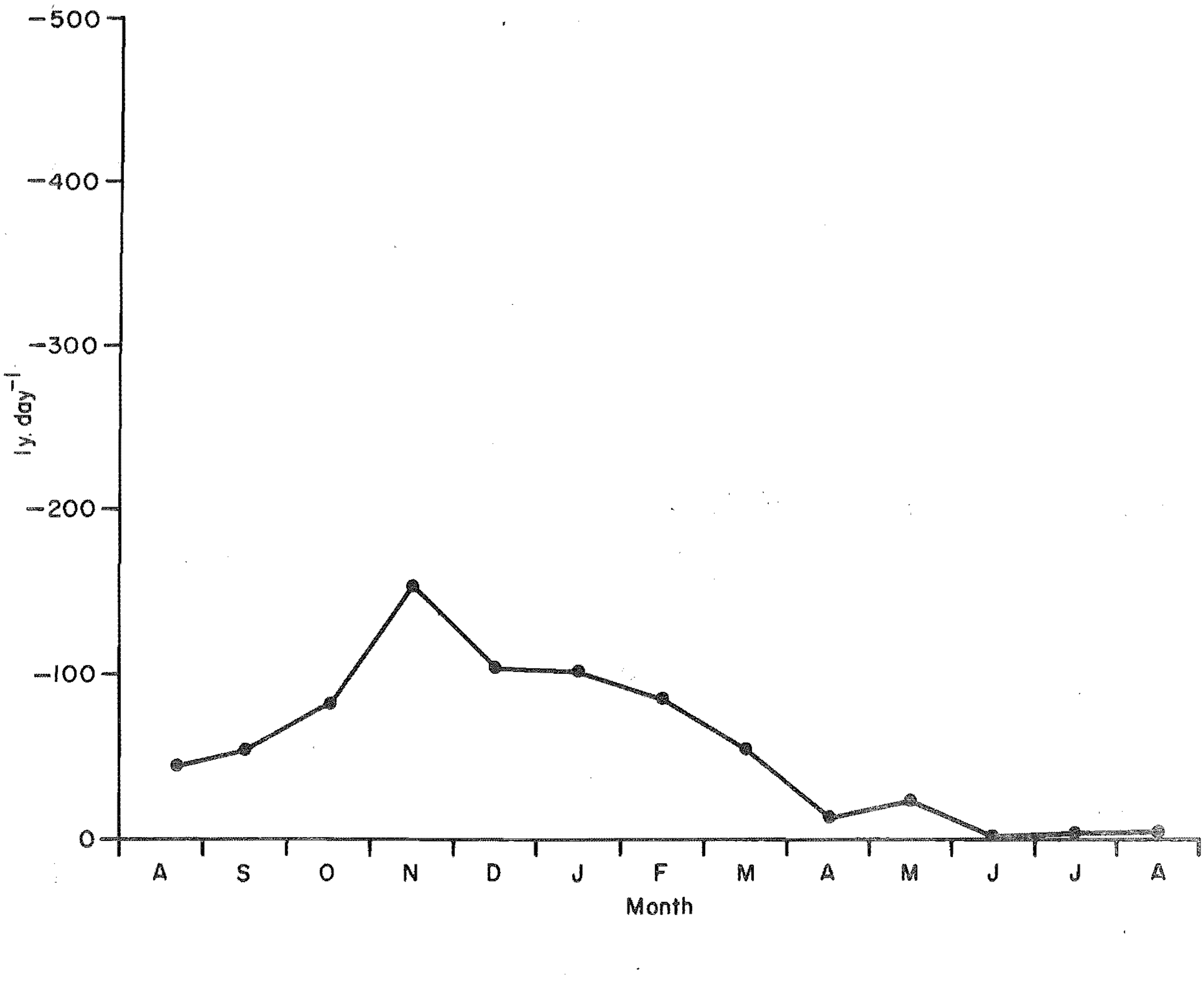


FIGURE 5.1: Monthly mean values of sensible heat transfer at the Chilton Valley during the study period

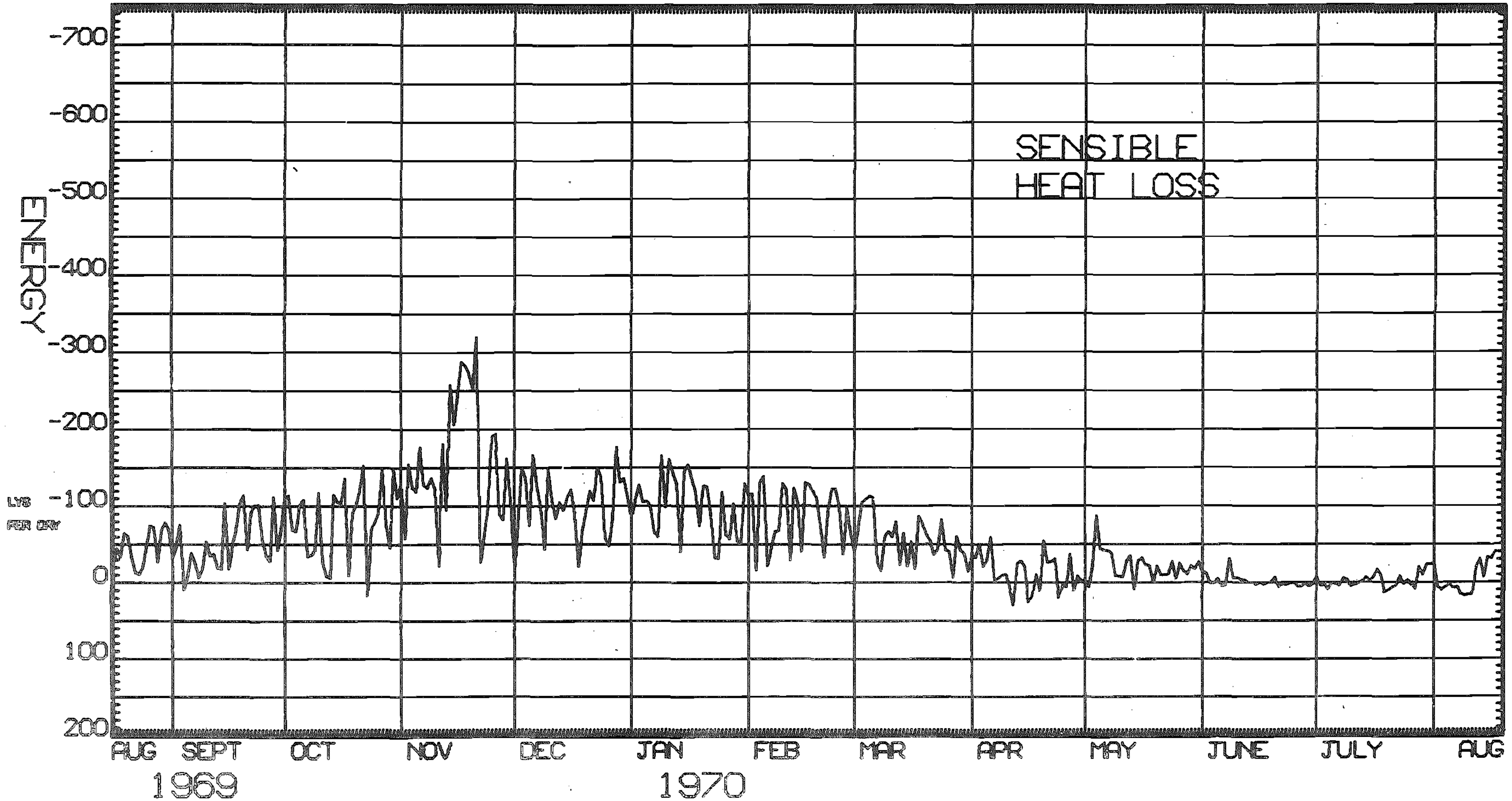


FIGURE 5.2: Daily values of sensible heat transfer at the Chilton Valley during the study period

Ontario, while Frankenberger (1962), in Germany, found a range of -1.51, in winter, to 0.86 in spring. Where there is a dry season Bowen ratios can rise as, for example, in south east India where a range of 0.6 to 4.3 has been recorded (Polavarapu, 1968). On the basis of these values, the Chilton Valley results were relatively high for the early part of the study year and the summer, and approached those of a dry climate in November. From April to August 1970, with the exception of May, values were low, indicating that most available heat (from all sources) was lost, mainly in evapotranspiration. However, the monthly mean values did not become negative as was found in Germany.

Daily values of P (Fig. 5.2) showed the high day to day variability that was apparent in the values of the other heat balance components. The high values of P during the dry period in November were outstanding, and a value of -320 ly day^{-1} was estimated at this time. In an environment where the growth of vegetation is imperative it is regrettable that part of this energy was not utilized in the evapotranspiration process. Relatively large positive flows (20-30 ly day^{-1}) occurred in April, and smaller positive flows occurred quite frequently in June, July, and August, 1970. In April the positive values of P helped to support high values of LE, but in the winter months downward flows of sensible heat helped to sustain either small LE values or negative Rn values or both (see Appendix F).

5.3 Daily Wind Values

Wind velocities were obtained by analysis of the Lambrecht anemometer chart and taking the average velocity for each hour (to the nearest m sec^{-1}) and then computing the daily wind velocity. These values were checked by using hand anemometers and a totalizing cup anemometer. Comparison between the last two showed that the totalizing anemometer underestimated wind speeds by 35% at 2.8 m sec^{-1}, with this percentage decreasing with wind speed. The possibility of underestimation by the latter had previously been communicated to the author by its owners, Department of Botany, University of Canterbury (M. Sinclair pers.comm.). The hand anemometers were assumed to be the least inaccurate of the three instruments. When the Lambrecht records were being calibrated against the totalizing anemometer, the values given by the latter were increased by 35% if the mean wind speed recorded was greater or equal to 3.0 m sec^{-1}. Even having allowed for this, values from the Lambrecht recorder were 25% larger at 5 m sec^{-1} and 40% larger at 8.5 m sec^{-1} than those of the totalizing anemometer. This is thought to be due to the method of data extraction from the original chart. An independent test showed that the person extracting average hourly values was attracted more by the higher than the lower value wind velocity traces. Comparison with 288 values of daily wind run totals between the Chilton Valley and the Craigieburn Forest Service station also showed the Valley records to be higher by an average of 27%. Thus, to make allowance for this, daily totals taken from the Lambrecht chart were multiplied by a factor of 0.73. In absolute terms this has a greater effect on the higher daily totals than the lower ones, which is in accordance with the possible errors due to data extraction. Owing to these uncertainties it was impossible to obtain a direct estimate of the accuracy of daily wind values. However, an indirect estimate was made by computing the relationships between (1) the adjusted Lambrecht records and the totalizing anemometer records (which had been adjusted for their underestimation) for the period January 20 - February 19, 1970, and (2) between the adjusted Lambrecht records and the Craigieburn records which were used to complete values for days of missing Lambrecht data. In the absence of any direct method of estimation of accuracy, the S.E.E.of these relations was used to give an idea of the accuracy of the daily Lambrecht totals. This was $\pm$ 0.7 m sec^{-1} or, where Lambrecht records were absent, $\pm$ 1.2 m sec^{-1}.

The mean wind velocity for the study year was 2.4 m sec^{-1}. Monthly variation of daily wind velocities (Table 5.2) showed higher velocities to have occurred from July to November, and lower velocities from December to March. At the Camp Stream site in the Craigieburn Range during 1961-63, winter and spring were also found to be the windiest seasons, and the annual average velocity was 2.5 m sec^{-1}. The annual wind velocity in the study period was, as expected owing to greater shelter, lower than the value recorded in earlier years at the Biological Station (section 1.4). The seasons of the highest average wind velocities,however, were the same at the two locations.

An analysis was made of individual hourly values of wind speed and direction at the Chilton Valley site (Table 5.3). The totals and percentages of frequency of wind direction confirmed that the prevailing wind was a down-valley wind from

TABLE 5.3

ANALYSIS OF HOURLY VALUES OF WIND SPEED AND DIRECTION AT THE CHILTON VALLEY DURING THE STUDY YEAR. VELOCITY VALUES IN m sec^{-1}

	Absent	N	NE	E	SE	S	SW	W	NW
September									
Calm	22								
1 - 5	344								
6 - 10	338	No direction data available							
11 - 15	16								
16 - 20	0								
October									
Calm	17								
1 - 5	316								
6 - 10	380	No direction data available							
11 - 15	31								
16 - 20	0								
November									
Calm	9								
1 - 5	211								
6 - 10	138								
11 - 15	2	No direction data available							
16 - 20	0								
December									
Calm	203								
1 - 5	0	74	36	7	5	145	27	11	5
6 - 10	0	3	20	4	1	21	6	3	4
11 - 15	0	0	1	0	0	0	0	0	0
16 - 20	0	0	0	0	0	0	0	0	0
January									
Calm	124								
1 - 5	0	183	182	37	24	86	3	3	8
6 - 10	0	12	56	4	7	12	0	0	1
11 - 15	0	0	0	0	2	0	0	0	0
16 - 20	0	0	0	0	0	0	0	0	0

Table 5.3 (Contd.)

	Absent	N	NE	E	SE	S	SW	W	NW
February									
Calm	248								
1 - 5	0	111	66	27	20	65	19	14	11
6 - 10	0	31	30	11	0	4	2	1	2
11 - 15	0	2	6	2	0	0	0	0	0
16 - 20	0	0	0	0	0	0	0	0	0
March									
Calm	473								
1 - 5	0	78	42	21	12	61	3	9	9
6 - 10	0	14	10	4	2	4	0	1	0
11 - 15	0	0	0	0	0	0	1	0	0
16 - 20	0	0	0	0	0	0	0	0	0
April									
Calm	115								
1 - 5	0	180	68	28	4	27	3	6	12
6 - 10	0	77	45	11	0	3	0	0	0
11 - 15	0	7	10	1	0	0	0	0	0
16 - 20	0	1	2	0	0	0	0	0	0
May									
Calm	29								
1 - 5	0	121	24	9	3	32	28	23	24
6 - 10	0	15	2	0	1	8	10	2	1
11 - 15	0	0	0	0	0	2	2	0	0
16 - 20	0	0	0	0	0	0	0	0	0
June									
Calm	39								
1 - 5	0	309	80	38	10	44	24	21	7
6 - 10	0	33	10	0	0	3	1	2	1
11 - 15	0	0	1	0	0	0	0	1	0
16 - 20	0	0	0	0	0	0	0	0	0

Table 5.3 (Contd.)

	Absent	N	NE	E	SE	S	SW	W	NW
July									
Calm	75								
1 - 5	0	268	119	41	3	37	23	22	2
6 - 10	0	57	48	19	0	2	4	1	0
11 - 15	0	10	11	0	0	0	0	0	0
16 - 20	0	2	0	0	0	0	0	0	0
August									
Calm	19								
1 - 5	0	97	55	11	2	14	19	18	11
6 - 10	0	28	49	13	0	2	6	2	1
11 - 15	0	4	6	0	0	0	0	0	0
16 - 20	0	1	0	0	0	0	0	0	0
Year									
Calm	3152								
1 - 5		1418	672	219	83	511	149	127	89
6 - 10		270	270	66	11	59	29	12	10
11 - 15		23	35	3	2	2	3	1	0
16 - 20		4	2	0	0	0	0	0	0
% of all Observations	43.6	23.7	13.6	4.0	1.3	7.9	2.5	1.9	1.4
% of obs. of wind		42.1	24.1	7.1	2.4	14.1	4.4	3.4	2.4

a northerly quarter. As has been mentioned before, the prevailing north west wind in the Cass Basin is altered by topography, to be a northerly or north easterly wind in the valley itself. The second most frequent wind came from the south, where the valley is open to larger scale southerly flow. The velocity categories indicate that not only was the wind from the northerly quarter the most frequent, but it was also the strongest. Examination of the monthly analyses showed that December and March were months of relatively few winds from the northerly quarter, and Table 5.2 indicates that these months had relatively low average wind velocities. The highest wind gust in the study period was one of 27.8 m sec^{-1} (62 m.p.h.) recorded on 7 September 1969 and again on 26 October 1969. The relatively sheltered position of the valley recording site is reflected in the high percentage of calms recorded in the hourly observations. Examination of the original recorder charts showed that most of the calm periods occurred at night as might be expected.

5.4 Wind Profiles and Wind Field Studies

A pilot study of wind profiles and the spatial variation of wind was made on 4 and 5 October 1969. The profile data were used to examine the applicability of aspects of aerodynamic theory and empiricism in the present location. The study of the spatial variation of wind, indicated the areas of the valley where values of turbulent transfers might be expected to be greatest, and the extent to which wind velocities, measured at the recording site, were representative of the valley as a whole.

Three hand anemometers were used. Wind measurements taken by these, at the same location and time, had standard deviations of 0.025 m sec^{-1} and 0.06 m sec^{-1} at average wind speeds of 2.17 and 2.84 m sec^{-1} respectively. Measurements of wind profiles were taken at several points in the valley (Fig. 5.3). The mean of five, or in some cases ten, half minute averages was taken for one profile (Table 5.4).

In an examination of the profile data (Fig. 5.4), profiles C and F may be immediately singled out as being anomalous, since they indicate inversion conditions which did not in fact apply at the time of measurement. Both profiles were taken in positions near to high bush and this may be a possible cause of the anomaly. All of the other profiles, except E, show some convexity toward the velocity axis which is to be expected under unstable lapse conditions that existed at the time of measurement (Sutton, 1954, p.234). Profile G was taken in almost the same place as profile E. The former was measured under conditions of greater instability and was computed from more readings than that of E. It is therefore considered more reliable.

Despite factors mentioned earlier (section 5.1), such as the inhomogeneous nature of the surface, and, in addition, the non-neutral lapse conditions existing at the time, an analysis of roughness length, the exponent in the power law, and the friction velocity, was made. Data from profiles C, E, and F were excluded from the analysis owing to the apparent anomalies displayed by these profiles in the light of lapse conditions. The roughness length z_o is taken to be the intercept of the wind profile on the log z axis in Fig. 5.4. The empirically

TABLE 5.4

WIND PROFILE DATA

Profile Position	Surface Cover	Slope Angle Degrees	Horizontal Wind Velocities (m sec^{-1}) at heights (m) Shown in Brackets	Av. Lapse Rate $^{o}C/100$ m
A	Bare scree	25 - 30	3.20 (0.31), 4.63(0.91), 5.42(2.44)	- 7.6
B	Shortgrass, Tussock, Cassinia and Matagouri	25 - 30	1.22(0.31), 4.60(0.91), 5.95(2.44)	- 7.6
C	Manuka $\geq$3 m high. Taken in partial clearing downwind of bush.	25 - 30	0.16(0.31), 1.42(1.82), 3.20(2.74)	- 7.6
D	Manuka ~ 1.5 m high	15 - 24	1.22(0.62), 3.62(1.52), 4.78(2.58)	- 7.6
E	Tussock and scrub near recording huts	0 - 14	2.08(0.31), 4.40(1.52), 5.08(1.98)	- 7.6
F *	Bare scree	25 - 30	4.53(0.31), 5.53(0.91), 7.73(2.44)	- 5.3
G *	Tussock and scrub near recording huts	0 - 14	1.94(0.31), 3.50(0.91), 4.42(2.44)	- 5.3

* The average of ten readings were taken for these profiles.

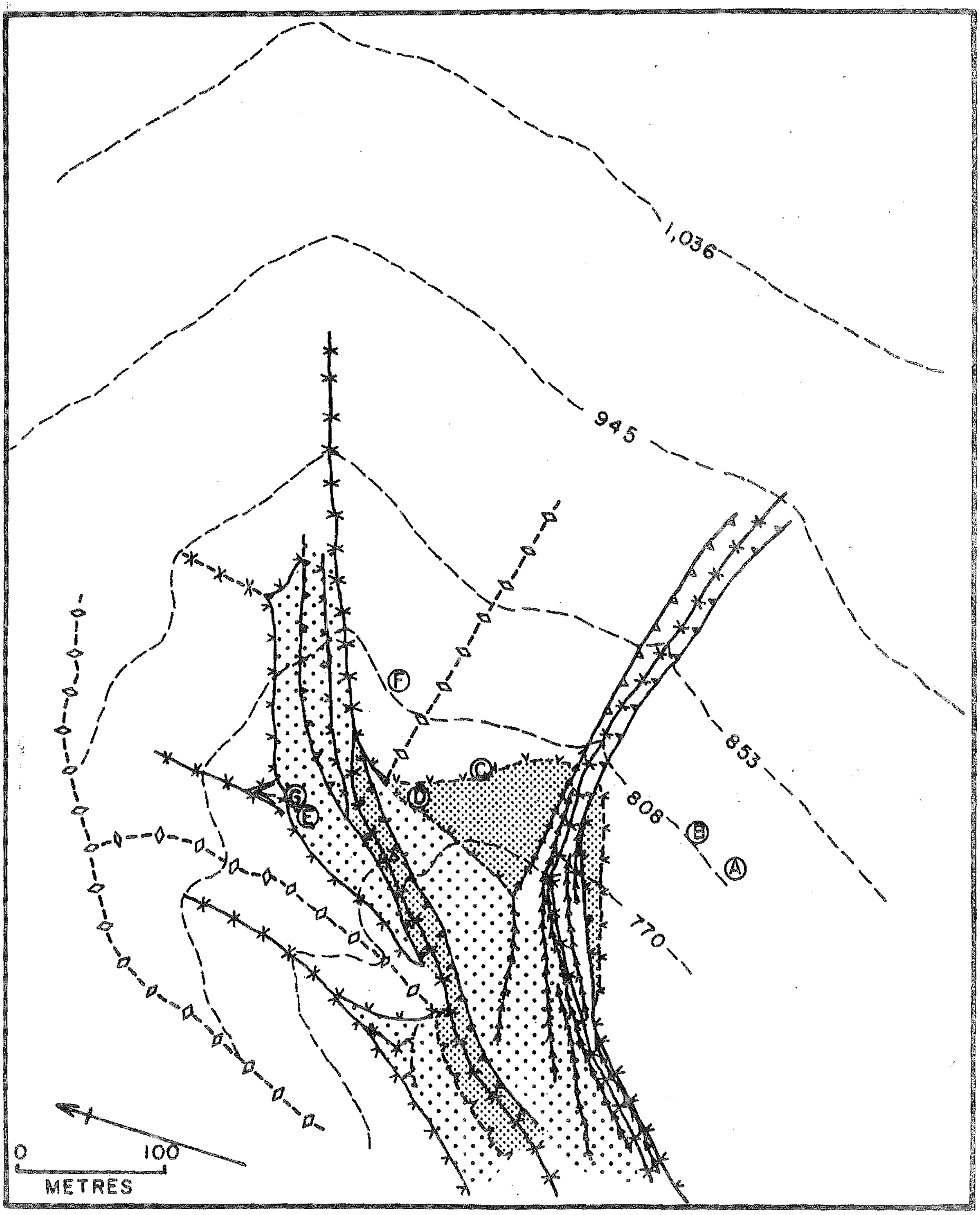


FIGURE 5.3: Location of positions where wind profiles were taken on 4 and 5 October 1969

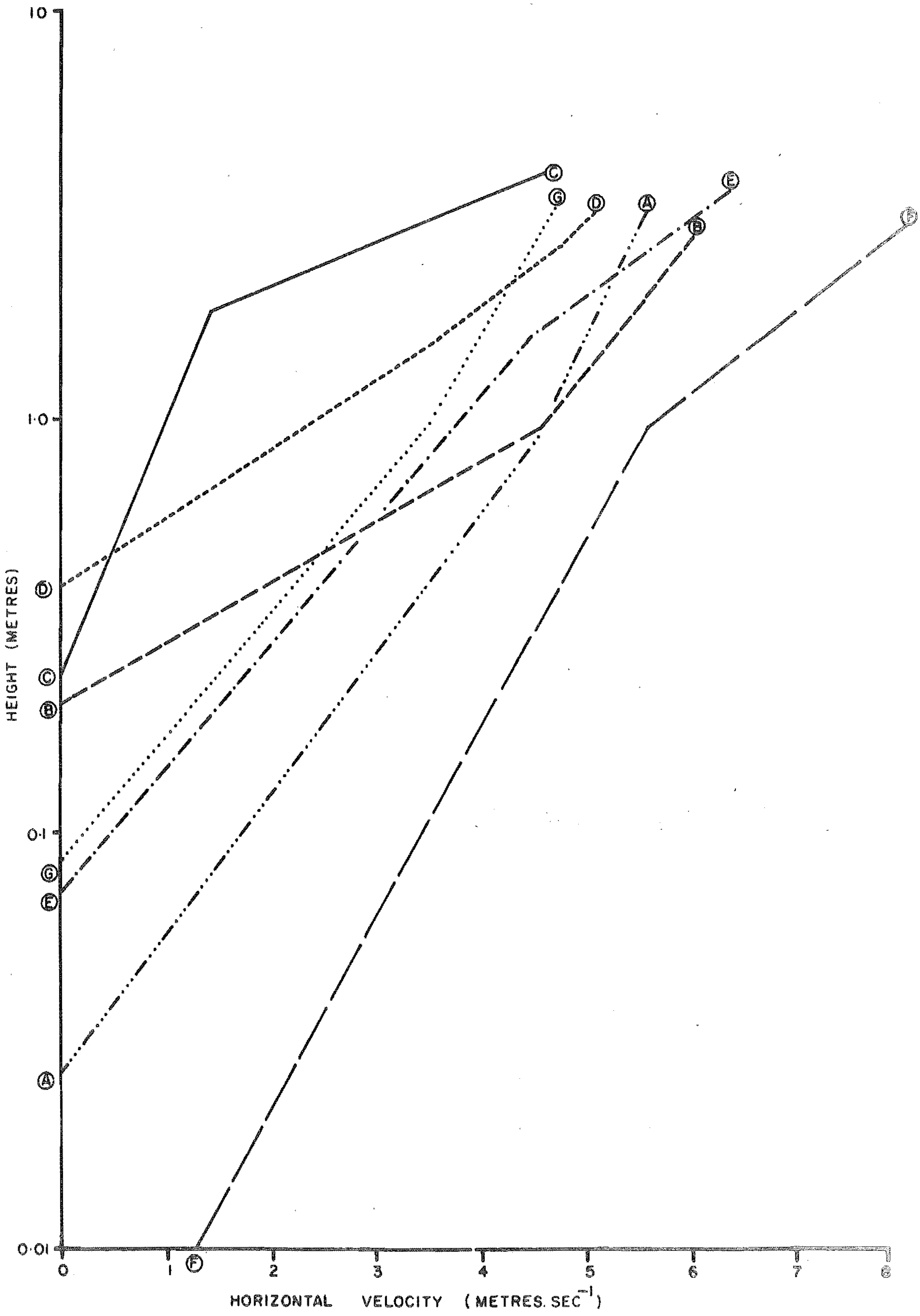


FIGURE 5.4: Values of wind profiles taken on 4 and 5 October 1969

derived power law expressing the wind velocity, u, at some height, z, as a function of the wind velocity at some reference level, say one metre, u_1, and some exponent, a, may be written as

$$u = u_1 \ (z/z_1)^a \qquad 5.4.1.$$

A theoretically based method of expressing the wind velocity change with height is the Prandtl equation

$$u = \frac{u_*}{k} \ln \ (z/z_o) \qquad 5.4.2.$$

from which the value of u_*, the friction velocity, can be computed if the values of the other terms are known.

In general the values of roughness lengths (Table 5.5) compare favourably with those in summaries of values of this parameter quoted by Sutton (1953, p.223) and Sellers (1965, p.150). More specifically, their validity can be tested by applying a relationship of Kung and Lettau, given by Tanner and Pelton (1960), between values of roughness length and vegetation height. Using this relationship, the roughness lengths in Table 5.5 would correspond to vegetation heights of 20, 163, 311 and 62 cm respectively for profiles A, B, D and G. On the basis of these results it would appear that the higher vegetation found in the valley played a more dominant role in determining roughness lengths than did the lower vegetation. The result for profile D however, which indicated a vegetation of over 3 m was abnormally high, even though the profile site had vegetation of this height about 40 m upwind of the position.

The values of the power a, in equation 5.4.1., all appear to be high in comparison with those found elsewhere. Henning,

TABLE 5.5

VALUES OF PARAMETERS DERIVED FROM WIND PROFILE MEASUREMENTS

	PROFILES			
	A	B	D	G
Approx. height of Vegetation cms	0	10-150	150	5-150
Roughness length z_o cms	2.7	21.0	40.0	8.0
Power of Power Law (a)	0.40	1.39	1.02	0.39
Friction Velocity u_* m sec^{-1}	0.52	1.19	1.10	0.57
Velocity at 2 m m sec^{-1} $X10^{-1}$	0.53	0.57	0.43	0.47

reported by Geiger (1965, p.117), found a seasonal variation in a of 0.28 to 0.53. An even wider range of values has been reported, from almost zero under very unstable conditions (Ali, 1932) to 0.85 under very stable conditions (Best, 1935). But even when the known decrease of a with height (Molga, 1962) is taken into account, the values in Table 5.5, especially those for profiles B and D, are still unusually high. These results suggest therefore that the particular conditions existing in the Chilton Valley had a marked effect. It is possible that if wind velocities were observed at levels higher than 3 m, the influence of the surface vegetation could be determined, and some zero displacement term applied. If these conditions were fulfilled more reasonable values of a might be found.

The effect of surface conditions in the Chilton Valley is also seen when the friction velocities are considered. These were computed from equation 5.4.2. Data of Covey et.al. (1958) for short grass (z_o = 0.75 cm) in Nebraska show a variation in u_* from 0.23 to 0.40 m sec^{-1} in different stability conditions during the day. Sutton (1953 p.233) reports a variation from 0.16 m sec^{-1}, for very smooth surfaces, to 0.63 m sec^{-1}, for thick grass up to 50 cm high. He has also suggested that u_* equals approximately a tenth of the wind speed at 2 m. Although Sellers (1965), referring specifically to short grass, has quoted the same fraction but with respect to velocities at 0.4 m, Sutton's generalisation is used here for comparative purposes, (Table 5.5). Using Sutton's criterion it can be seen that the values of u_* for profiles A and G are reasonable but those of profiles B and D are more than twice as high as would be expected.

Conclusions from this preliminary study must await more data for confirmation, but it appears from the above discussion that, in general, the presence of a large variety of vegetation of differing heights precludes the measurement of useful wind profile data by means of the simple methods that have been employed. This is particularly true of the north west facing slope, where stands of high (3 m) vegetation are mixed with bare scree slopes. Also, judged by the values of a that were found, the commonly applied power law should be used with caution in the lower 3 m at this location. On the other hand, on a more tentative basis, the evidence from profile G suggests that approximate values of wind profile parameters can be obtained, using the present methods, for the area at the floor of the valley, near the recording site. Despite the 'stability' indicated by profile E which was taken in a similar position, it is encouraging that the roughness lengths of both profiles were similar and further studies of this type are warranted for this location. In any re-measurement of the profiles in the positions reported here, wind observations to greater heights would be advantageous.

On 5 October a study was made of the spatial distribution of wind velocities in the valley. The day of the study had an average wind velocity of 4.6 m sec^{-1} as recorded by the Lambrecht anemometer. During the whole day the wind blew down the valley from the north and north east. Wind velocities were measured with three hand anemometers along two lines across the valley (Fig. 5.5). One anemometer was kept at the same position in the valley floor to act as a control, so that other readings, along the line, could be adjusted for overall wind velocity changes with time. All results (Table 5.6) are the mean of ten, half minute readings.

The results of the cross valley wind velocity observations indicated the north west facing slope to have been slightly windier than the south east facing slope. An area north east of the position of the recording huts (i.e. positions 2 and 3 on line A) also had high velocities. This may have been due to channelling of air through a saddle in the ridge on the north west side of the valley. A low wind velocity was recorded at point 4 on line B, which was in the small, sheltered gully in the centre of the valley.

Wind velocities were also measured along the length of the valley at positions 8, 9, 10 and 11 (Fig. 5.5). The results (Fig. 5.6) showed an increase in velocity with distance up the valley. The maximum velocity of the ten readings taken was also largest at the higher altitude. These results are in agreement with those found in other areas of the High Country (Coulter, 1967).

Apart from the interest that these observations on the wind field of the valley have from the general climatic viewpoint, they are relevant to energy exchange studies in the following way. Bearing in mind that the downvalley wind is the predominant one in the valley (see section 5.3) it is seen that the location of the recording site is fairly representative of wind conditions in the valley as a whole. However, owing to increased wind speeds, processes of turbulent energy exchange may be expected to occur at higher rates on the north west facing slopes and at lower rates on south east facing slopes. From general observations it may

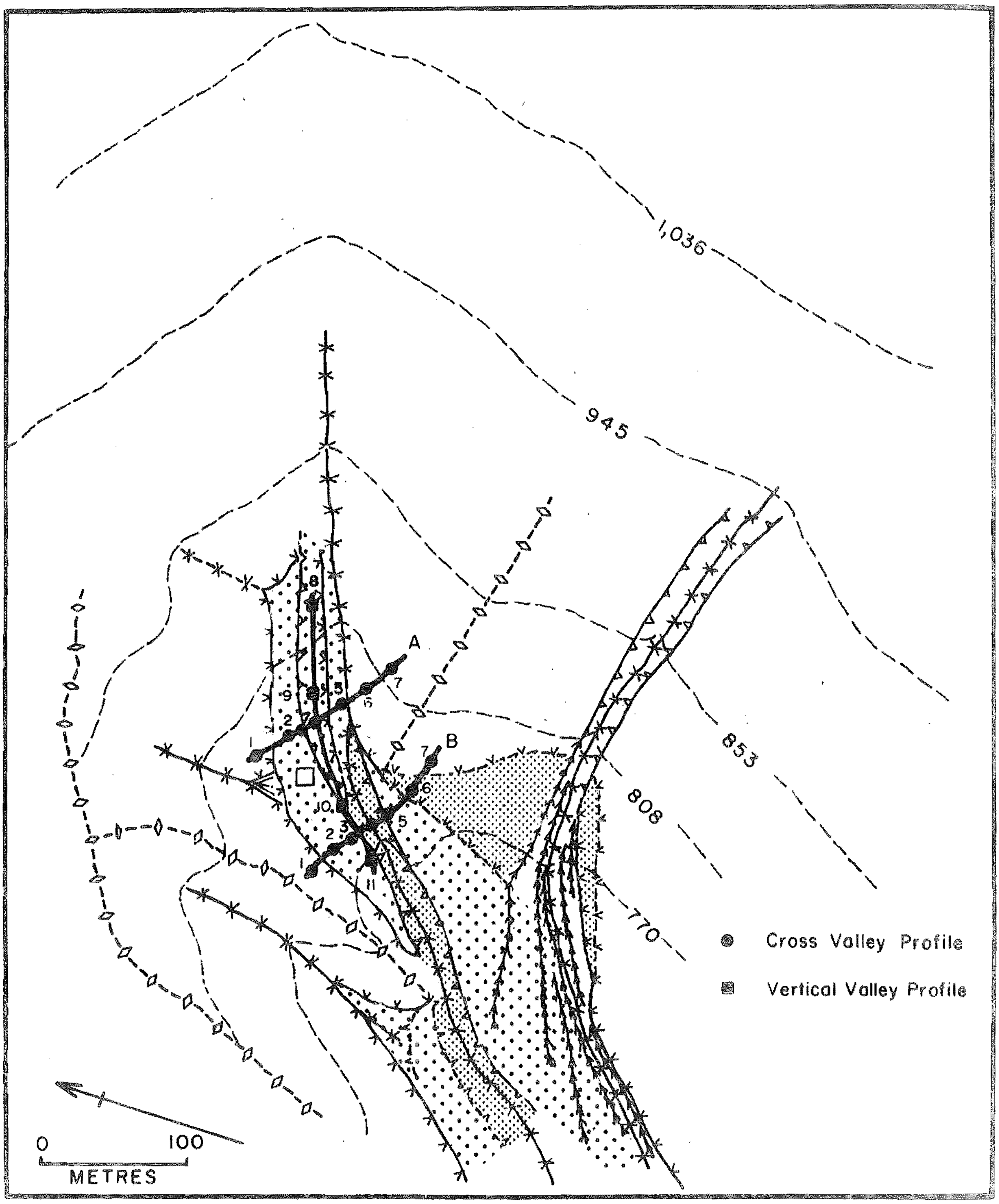


FIGURE 5.5: Location of positions where wind field observations were made on 4 and 5 October 1969

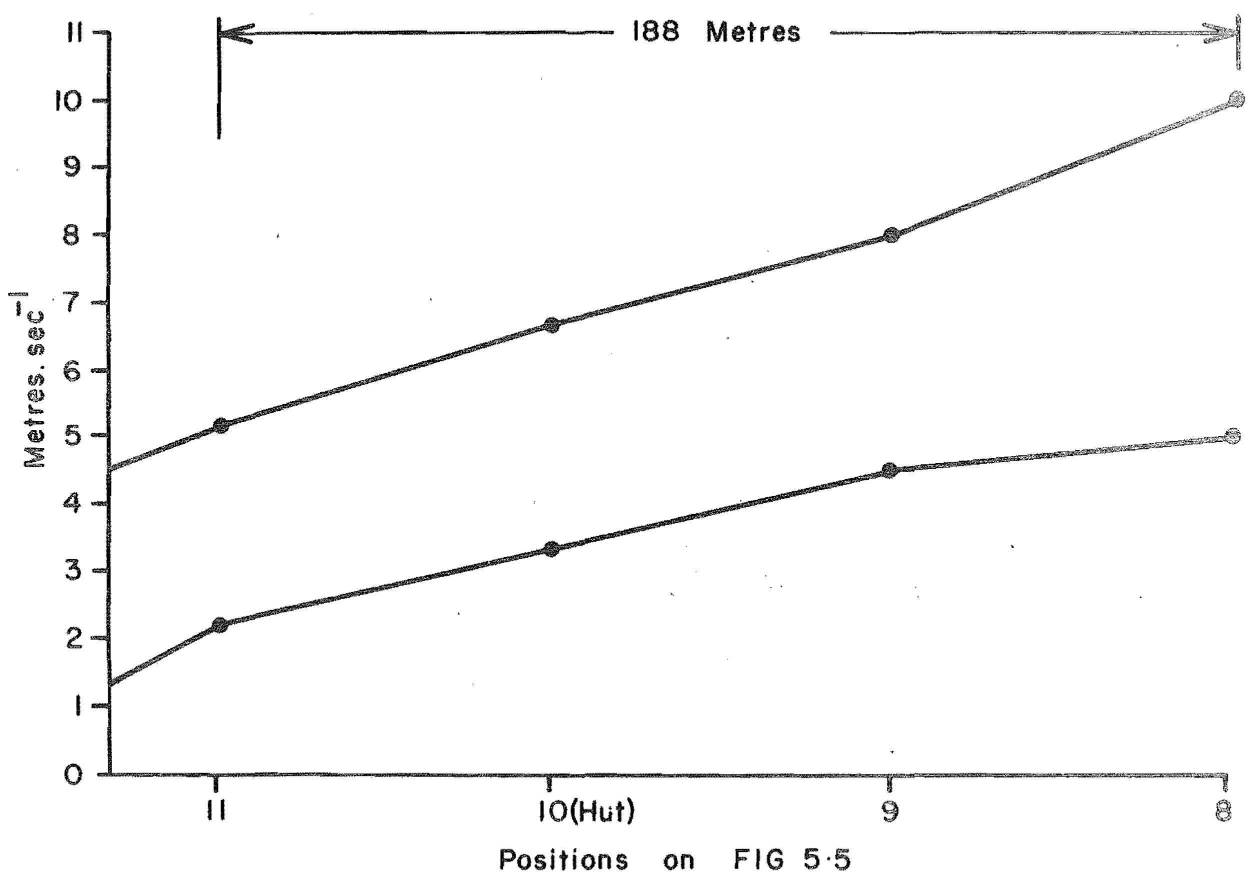


FIGURE 5.6: Values of wind velocity at 2 m in a valley length profile observed on 4 and 5 October 1969. Values of the maximum (top line) and mean (bottom line) of ten half minute averages, are shown

TABLE 5.6

WIND VELOCITIES AT POINTS ON LINES ACROSS THE VALLEY. VELOCITIES IN m sec^{-1}

Position	1	2	3	4	5	6	7
Line A	2.4	4.5	4.5	4.0	4.1	3.5	3.2
Line B	3.6	2.5	4.1	2.2	5.4	3.6	4.0

TABLE 5.7

NUMBER OF DAYS WITH VALLEY INVERSIONS OCCURRING PER MONTH IN THE CHILTON VALLEY

August	11
September	8
October	4
November	21
December	7
January	9
February	10
March	9
April	11
May	23
June	15
July	4
August	-

be speculated that the reverse is true under southerly conditions, although in both cases, wind velocities would increase with height. Clearly, further work is needed to substantiate these points.

5.5 Advection

Advected heat can affect the heat balance by changing the values of the heat balance components. A complete resolution of the effect of advective heat on each of the individual energy balance components is impossible with the present data. However, there are several approaches that may be used to indicate the frequency, as opposed to the magnitude, of advective effects.

In the present situation advection occurs on two scales. Firstly, there is the relatively small scale and least important effect concerned with diurnally alternating mountain and valley winds. Secondly, there is advection associated with larger scale airflows. In particular, north westerly and southerly air flows are important in the present location. The frequency of advection on both small and large scales will be examined, and the effect of large scale advected heat on the evaporative and sensible heat flow terms will be investigated as far as possible.

Observation has shown that mountain and valley winds certainly occur in the Chilton Valley, but that both of them, and especially the day time upvalley wind, are often overshadowed by the larger scale airflows. This occurs because the alignment of the valley is particularly conducive to channelling the predominant large scale north westerly and southerly winds along the north east-south west axis of the valley. However, an idea of the frequency of down valley winds due to cold air drainage can be gained by examining the number of days with valley temperature inversions, since these occur when the effect of the larger scale winds is minimal. 'Valley temperature inversions' are so termed here to distinguish them from the 'true inversions' (i.e. the occurrence of increased temperature with elevation at one site) that were briefly examined in section 1.4. For the purposes of this investigation, and owing to the availability of data extracted from the original charts, a day on which a valley inversion occurred is defined as one where the minimum air temperature at the hill site was greater than the minimum temperature at the recorder site. This gives only an approximate guide since it assumes that the minimum temperatures occurred at the same time, which is not always true.

Valley inversions (Table 5.7) were most frequent in November, March and June. Throughout the study period they occurred on 38% of the days sampled. The valley inversions were almost always accompanied by true inversions, as measured between the 3 m and 12 m mast thermistors. In order to examine the size of P at these times, the empirical data of Frankenberger (1962) were used. If the katabatic wind speed was 2 m sec^{-1} and the temperature gradient between 3 m and 12 m was $0.5^{o}C$, to take a rather extreme example, Frankenberger's data would show a value of P of less than +0.8 ly hr^{-1}. Since the wind speed often decreases to zero on nights of inversions, taking the above rate of heat flow for a period of only six hours would result in a downward flow of sensible heat of not more than 5 ly. This is usually small compared with the daily totals of P and those of the other heat balance components. The application of Frankenberger's data gives only an approximate guide as it is assumed that the exchange coefficients at his site and those in the Chilton Valley are the same. However, on the basis of the above discussion it would appear that the effect of advected air which is associated with temperature inversions and down valley winds, on the sensible heat flow is small but not necessarily insignificant.

The larger scale effect of the föhn wind is probably much more important with regard to advection. As an example, the work of Lamb (1970), who studied the effect of north west winds on the Canterbury Plains in summer, may be cited. The föhn wind exercises its most extreme influence on the Plains. Lamb found values of downward sensible heat flow averaging 45 ly day^{-1} and rising to 113 ly day^{-1}. These values may be overestimates owing to the manner by which they were computed, but it is clear that the föhn conditions gave rise to considerable amounts of advected heat.

In the case of the Chilton Valley, an example has already been given (section 3.5) of north west conditions leading to a value of P of 29 ly day^{-1}. Moreover, the first Penman model indicated conditions, especially in winter, demanding considerable downward flows of sensible heat (Table 5.8). In this table a similar pattern is shown by the results of the modified Penman model but the size of the flows is not so large. An analysis (Table 5.8) was also made of the number of days when the 'drying power' term accounted for more evaporation than the radiation term in the Penman model. This

TABLE 5.8

EVIDENCE OF THE FREQUENCY AND EFFECT OF ADVECTION DURING THE STUDY PERIOD

	Largest downward flow of sensible heat given by the original Penman Model ly day^{-1}	Largest downward flow of sensible heat given by the Modified Penman Model ly day^{-1}	No. of days where drying power term exceeded the Radiation term in Modified Penman Model
August	+ 30	+ 9	0
September	+ 49	-	1
October	+113	+ 16	1
November	+ 33	-	1
December	-	-	0
January	+ 4	-	0
February	+ 12	-	0
March	+ 1	-	0
April	+ 26	+ 30	3
May	+ 44	+ 9	2
June	+ 74	+ 6	20
July	+167	+ 13	16
August	+139	+ 18	8

gives information on the times when advection had a large effect, relative to that of the radiant heat source, on the latent and, by implication, sensible heat fluxes.

The above results and discussion do not altogether give a clear picture of the effect of advection in summer, but they indicate that at least in winter, advection is frequent and has a relatively large effect on the sensible and latent heat flows.

A more rigorous analysis of the frequency of large scale advection may be made by examining the deviation of the diurnal temperature variation from its radiation-controlled cycle. Anomalous times of occurrence of maximum and minimum air temperatures during a period of a day, have been taken by Karapiperis (1953) to estimate the frequency of both warm and cold air advection. A similar analysis (Tables 5.9 and 5.10) for the Chilton Valley demonstrated the following important points.

1. The percentage of days with maximum air temperatures occurring between the expected times of 1200-1600 hr, 61%, was relatively low.

2. The percentage of days with minimum air temperatures occurring between the expected times of 0400-0800 hr, 39%, was very low.

3. Days when the temperature continually rose all through the day would give maxima between 2200-2400 hr, and these occurred on 3.5% of the period. Similarly, days when the temperature continued to fall all through the day would give maxima between 0000-0200 hrs and this occurred on 4.4% of the period. Both of these cases are days when the amplitude

TABLE 5.9

NUMBER OF DAYS DURING THE STUDY PERIOD WHEN MAXIMUM AIR TEMPERATURES OCCURRED BETWEEN SPECIFIED HOURS

	Time of Occurrence of Maximum Air Temperatures											
	0-2	2-4	4-6	6-8	8-10	10-12	12-14	14-16	16-18	18-20	20-22	22-24
August	0	1	0	0	0	0	7	9	0	0	0	0
September	3	0	1	0	2	6	13	2	0	1	1	1
October	1	1	0	0	0	3	15	7	2	0	1	1
November	1	0	0	0	3	12	6	6	2	0	0	0
December	0	0	0	0	3	4	15	6	2	0	0	1
January	1	0	0	0	1	4	5	12	6	0	0	2
February	0	1	0	0	1	3	8	9	3	1	0	2
March	4	0	0	0	3	5	4	10	2	2	1	0
April	1	0	0	0	0	5	14	8	1	0	0	1
May	1	2	0	0	0	2	20	2	2	1	0	1
June	3	2	0	0	0	6	11	2	3	2	1	0
July	1	0	2	0	0	4	14	5	1	0	0	4
August	0	0	0	1	0	2	9	3	1	0	0	0
Year	16	7	3	1	13	56	141	81	25	7	4	13

TABLE 5.10

NUMBER OF DAYS DURING THE STUDY PERIOD WHEN MINIMUM AIR TEMPERATURES OCCURRED BETWEEN SPECIFIED HOURS

	Time of Occurrence of Minimum Air Temperatures											
	0-2	2-4	4-6	6-8	8-10	10-12	12-14	14-16	16-18	18-20	20-22	22-24
August	2	0	3	6	0	0	0	0	0	1	1	4
September	5	3	6	4	0	0	0	1	1	0	2	8
October	2	3	8	4	0	0	0	0	1	0	2	11
November	5	5	6	0	0	0	0	0	1	1	3	9
December	3	5	7	5	0	0	0	0	1	1	1	8
January	4	1	8	7	0	0	0	0	0	0	2	9
February	0	3	14	2	0	1	0	1	0	0	1	6
March	3	4	6	5	1	0	1	0	1	1	1	8
April	4	0	6	7	1	0	0	0	0	1	5	6
May	2	1	6	10	1	0	0	0	0	1	1	9
June	5	3	2	3	1	0	0	0	0	0	3	13
July	2	1	4	6	4	0	1	0	0	2	1	10
August	1	1	3	5	2	0	0	0	1	0	0	3
Year	38	30	79	64	10	1	2	2	6	8	23	104

of the radiation-controlled temperature cycle is less than the amplitude or difference of the observed temperature variation, and is more likely than not due to advective influences. However, such occurrences should be more or less reflected by the frequency with which the minimum temperatures occurred between 2200-2400 hr and 0000-0200 hr respectively, but these frequencies were in fact much greater. A possible explanation for this is described below.

4. The large percentage of days when minimum temperatures occurred between 2000 hrs through midnight to 0400 hrs, 53%, may be in part due to cold air advection, but probably more important, it is a function of inversion occurrence. The explanation is demonstrated by temperature records taken at the Biological Station in the basin floor and the valley recorder site for the night of 17-18 February 1967 (Fig. 5.7). The result of cold air flowing down valley into the bottom of the basin may be seen. The values of both sets of temperatures decrease until 2300 hr, at which time the lower record continues to decrease but the valley record increases slightly. This may be due to relatively warm air, which has been displaced from the basin floor by the incoming cold air, being raised to the level of the recording hut. The situation of warm air, evident as a thermal belt (Dunbar, 1966), migrating up-slope until the valley, or in this case basin, is completely filled with cold air, has been previously recorded by Young (1921), Keen (1968) and others. In view of the frequency of temperature inversions, it is likely

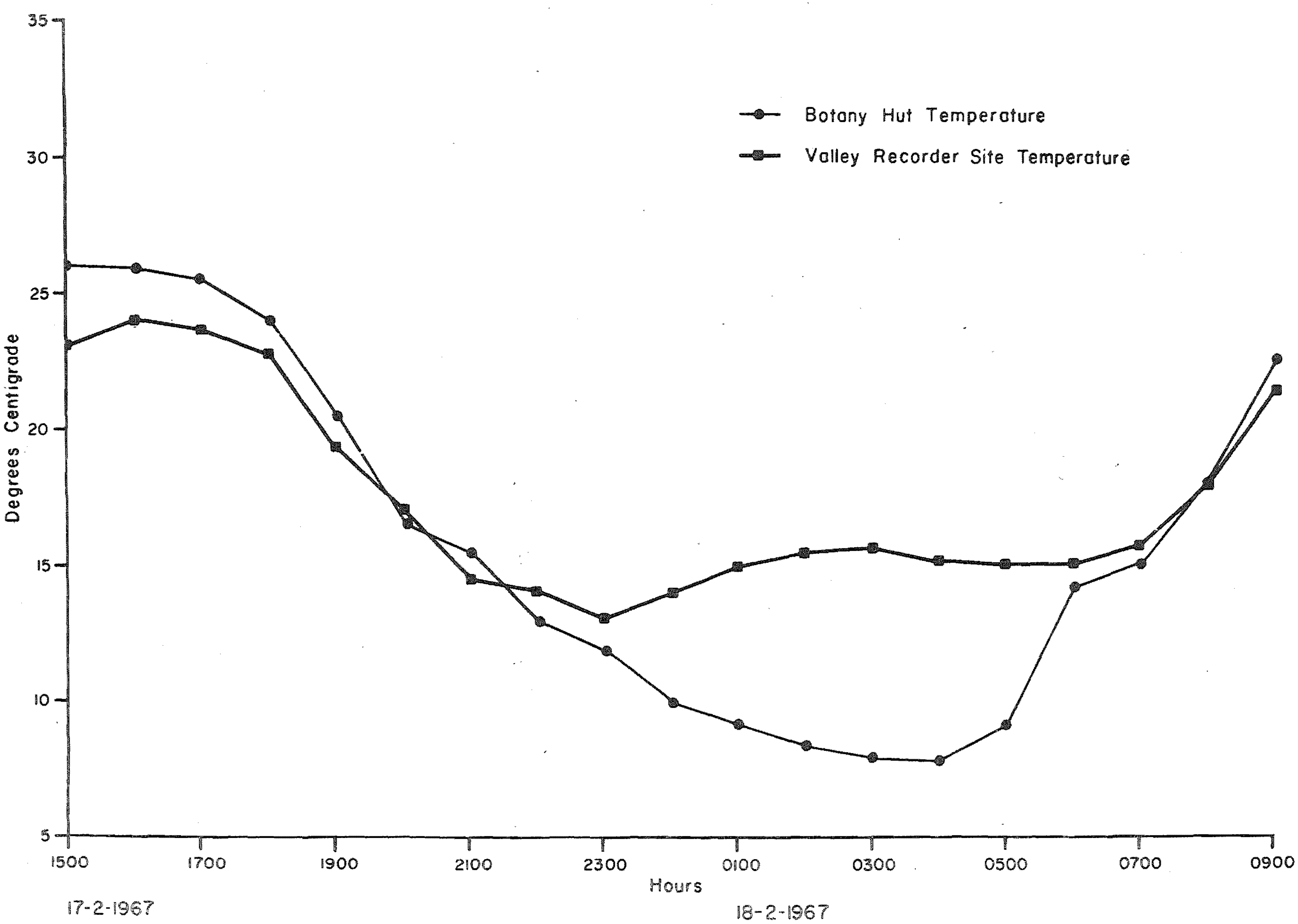


FIGURE 5.7: The development of a temperature inversion between the Botany Hut and the main recording site on 17 and 18 February 1967

to be common in the present location.

5. Ignoring the occasions when minima occurred between 2000 hr and 0400 hr described above, the percentage of other anomalous occurrences indicating cold air advection was 8%.

6. Neglecting the times when maximum temperature showed continual cooling or warming, the occurrence of anomalous maximum temperatures accounted for 32% of the period, although 15% and 7% of these times occurred between 1000-12000 and 1600-1800 hr respectively. However, there still remained a significant proportion of days showing warm air advection which was probably due to north westerly conditions.

7. There appeared to be no marked seasonal variation in the times of the cold and warm air advection described in 5 and 6, although the inversion situation was slightly more prevalent in the winter months.

The advection component for a climate can also be measured using the Aperiodic Advective Contribution component (AACC) developed by Flohn (1964). Here the dispersion of the mean air temperatures at each hour of the day, the periodic cycle (standard deviation = σ_p) which eliminates the random effects of advection, is compared with the dispersion of all the hourly temperatures (standard deviation = σ) which embodies advective influences. The periodic dispersion, σ_p, is small compared with the aperiodic, σ, in climates dominated by advective influences. The AACC, $(1 - \sigma_p/\sigma)$, cannot segregate advection that is continuous through the year such as land and sea breezes

TABLE 5.11

APERIODIC ADVECTIVE CONTRIBUTION COMPONENT FOR THE CHILTON VALLEY AND EDMONTON, CANADA

	σ_p	σ	σ_p/σ	$1-\sigma_p/\sigma$	$1-\sigma_p/\sigma$ For Edmonton Canada
August	2.36	3.43	0.69	0.31	
September	1.69	4.01	0.42	0.58	0.73
October	2.26	4.42	0.51	0.49	0.57
November	4.00	6.14	0.65	0.35	0.45
December	2.59	4.69	0.55	0.45	0.37
January	2.73	6.09	0.45	0.55	0.40
February	3.33	6.38	0.53	0.48	0.38
March	1.79	4.32	0.41	0.59	0.49
April	2.06	4.03	0.51	0.49	0.56
May	2.20	5.77	0.38	0.62	0.77
June	0.89	4.04	0.22	0.78	0.87
July	0.94	3.07	0.31	0.69	0.86
August	1.82	4.17	0.44	0.56	
Year	2.11	6.24	0.34	0.66	0.60

or presumably mountain and valley winds, but can be used to display regional differentiation. As an example the AACC is typically greater than 0.75 in middle and high latitude continents in winter and between 0.3 to 0.4 in summer. The high value in winter indicates the weakness of solar input. Table 5.11 shows the values of the AACC for the Chilton Valley recording site and for Edmonton, Canada (Flohn, 1964) a station which is sometimes subjected to outbreaks of cold polar air or warm tropical air. The comparison, using this technique, indicates the Chilton Valley to have a climate that is significantly affected by advective conditions.

These additional analyses therefore support the initial conclusion that there are times in the Chilton Valley when advected warm air can lead to downward flows of sensible heat, which on the basis of the results in Table 5.8, can be as great as 30 ly day^{-1}. This sensible heat may support greater evaporative heat losses than would otherwise have occurred. The effect is probably most marked in winter. There is evidence that cold air advection also occurs. The fact has emerged that cold air flow down the valley at night, while not having a large effect on the values of P and LE, may lead to the formation of a well pronounced thermal belt. If this is confirmed it would be of great biological and, indirectly, geomorphological importance.

5.6 Summary

The main conclusions and observations arising from this study of sensible heat flow and related phenomena are as follows:-

1. The value of P for the study year was -58 ly day^{-1}. Monthly values of P ranged from about -1 ly day^{-1} in winter, when flows on individual days were directed to the surface, to over 150 ly day^{-1} in November during times of soil moisture deficit. In December and January, when little or no soil moisture deficit occurred, sensible heat flow values were about -100 ly day^{-1}. Bowen ratios in summer were high compared with other mid latitude climates. This is possibly due to adaptations of the physiology of the vegetation to resist transpiration (see section 4.4.f.). There was a high day to day variability in sensible heat flow as has been seen in the other heat balance components.

2. Average wind speed in the valley in the study period was 2.4 m sec^{-1}. September and October had the highest average values. Over 66% of the valley wind came from the north or north east which is representative of the large scale north westerly wind. The north or north east wind directions were also associated with the highest wind velocities. Limited data indicated that the recording site is in a location fairly representative of the whole valley, although the spatial variation of wind in the valley is influenced by topography.

3. Limited data also suggest that it is possible to obtain some, useful values of parameters derived from the wind profile (e.g. the roughness length at the recorder site was estimated to be 7 - 10 cm). However, some aspects of empirical nature and aerodynamic theory are not necessarily applicable within the valley owing to the heterogeneous

qualities of the vegetation and surface. Larger numbers of observations, and observations to a greater height are necessary.

4. Advection of warm and cold air can affect the flows of sensible and latent heat. The climate of the Chilton Valley is significantly affected by advective conditions when compared with other mid latitude climates. Advection in the valley is associated with mountain and valley winds and large scale air flows, of which the north west föhn type wind is the most important. The advection of the large scale winds is believed to have a greater effect on the sensible and latent heat flows than the mountain and valley winds. The recorder site may be in a thermal belt created by cold air drainage at night.

CHAPTER SIX

PERIODICITY IN THE HEAT BALANCE COMPONENTS AND THE CLIMATONOMY OF THE CHILTON VALLEY

6.1 Introduction

The preceding chapters have been devoted primarily to the measurement or estimation of daily, monthly, and annual values of the surface heat balance of the Chilton Valley. The monthly and annual values and other relevant data for the study period are summarised in Table 6.1. In the present chapter, and in Chapter 7, an analysis of the values of the energy flows is presented. The flows themselves are a fundamental part of the climate of the Chilton Valley. Their relation to other aspects of the climate is also important. The succeeding analyses emphasise not only the importance and character of the energy fluxes, but also illuminate other notable features of the climate. The present chapter is concerned with periodicity in the values of the energy flows, and with the application of part of Lettau's (1968) theory of Climatonomy.

To a greater or lesser extent, periodicity is an important feature of most climates. In a thorough examination of the nature of any one climate, the role played by periodicity must

TABLE 6.1

VALUES OF HEAT BALANCE COMPONENTS AND OTHER PARAMETERS FOR THE CHILTON VALLEY DURING THE STUDY PERIOD

	Rn ly day^{-1}	LE ly day^{-1}	P ly day^{-1}	A ly day^{-1}	Mean daily Air Temperature T ^{o}C	Net Shortwave Radiation SW ly day^{-1}	Net Longwave Radiation LW ly day^{-1}
August	83	-48	-46	18	2.7	163	-80
September	112	-80	-54	22	6.9	208	-96
October	204	-114	-82	-1	6.3	349	-145
November	269	-102	-153	-13	12.8	449	-180
December	249	-135	-105	- 9	13.4	418	-169
January	265	-151	-101	-12	15.1	442	-177
February	224	-140	-87	3	13.1	380	-156
March	129	-77	-56	3	12.1	234	-105
April	97	-75	-12	7	10.0	184	-87
May	45	-29	-22	3	3.7	104	-59
June	9	-9	-1	2	3.9	48	-39
July	21	-25	-3	4	3.7	67	-46
August	55	-58	-4	7	6.9	120	-65
Year	136	-82	-58	2	8.8	244	-108

be specified. This may be achieved by means of time series analysis. In the first part of this chapter harmonic and spectral analyses are made of the values of the heat balance components, together with those of air temperature and SW↓. When the results of both analyses are viewed together (section 6.2.d.) the extent of periodicity in the climate of the Chilton Valley is apparent. Astronomically controlled cycles, and some due to other causes, are seen to have been present, and emphasise the dynamic nature of the character of the climate of the Chilton Valley.

In the second part of this chapter the annual cycle is also prominent in an application of Lettau's theory of climatonomy to the climate of Chilton Valley. The essence of this theory is that the climatic response in any location is determined by established physical laws and is predictable, given limited basic data. In section 6.3 climatonomic theory is used to estimate the annual values of the heat balance components and their variation through the study year. In the context of the present study the most important factor arising from the application of the theory is the revelation of firstly, the loss of water by percolation, and, secondly, the variability of the Bowen ratio, as being important features of the climate during the study year. However, the application also demonstrates the potential power of the theory and its use in making long term average estimates of the heat balance for the Chilton Valley and other High Country locations.

6.2 Time Series Analysis

6.2.a. Harmonic Analysis of the Daily Heat Balance Components

The technique of harmonic analysis as described by Conrad and Pollak (1962) was used with the aid of a computer programme for Fourier analysis written by Ralston and Wilf (1960). Daily values of the four main heat balance components were analysed and the explanation of variance of the original series by the first ten harmonics is shown in Table 6.2. The equation of the first harmonic for each component is presented in Table 6.3. An attempt was made to study the higher frequency cycles more closely by extracting the first harmonic from the original time series. The resulting explanations of variance are given in Table 6.4.

It can be seen that the annual cycles dominated almost all of the time series, and usually explained over half of the variance of the original series. When the first harmonic was extracted from the original series, despite the expected increased explanation of variance of the higher harmonics no one harmonic was particularly outstanding. However the fourth harmonic showed slightly higher explanation in the Rn, LE and P series. The results of the analysis of the A series are dissimilar from those of the other owing to the importance of the second harmonic. A full discussion of the results of these harmonic analyses is delayed until section 6.2.d., where they can be considered in relation to the results of the spectral analyses, and the possible physical causes of periodicity.

6.2.b. Spectral Analysis of Values of Daily Heat Balance Components

Daily values of the heat balance components were subjected

TABLE 6.2

RESULTS OF INITIAL HARMONIC ANALYSIS OF DAILY VALUES OF HEAT BALANCE COMPONENTS

Harmonic Number	No. of Days in One Cycle	% Variance Explained Rn	A	LE	P
1	365.00	64.29	28.53	50.49	55.43
2	182.50	0.20	30.05	1.31	1.31
3	91.25	0.24	4.96	1.02	0.58
4	45.63	1.28	4.97	2.22	2.38
5	22.82	0.01	4.50	0.32	0.36
6	11.41	0.44	0.88	1.87	1.77
7	5.71	0.70	0.76	0.58	0.97
8	2.86	0.23	0.05	0.39	1.70
9	1.43	0.02	0.18	0.83	0.60
10	0.72	0.21	0.29	0.80	0.18
Total Explanation by first 10 harmonics		67.62	75.17	59.83	65.28

TABLE 6.3

EQUATION OF THE FIRST HARMONIC FROM THE HARMONIC ANALYSIS OF DAILY VALUES OF THE HEAT BALANCE COMPONENTS FOR THE STUDY YEAR. UNITS OF THE EQUATIONS ARE LY DAY^{-1}. θ IS IN DEGREES OVER A BASIC INTERVAL OF 360°. 1° IS EQUIVALENT TO 1.0138 DAYS

Date of Max.	Date of Min.	Equation of First Harmonic	Equation No.
9/1/70	10/7/70	$Rn = 138.4 + 131.52 \sin(\theta + 326^{\circ})$	6.2.1.
20/12/69	20/6/70	$A = 1.6 + 8.78 \sin(\theta + 306^{\circ})$	6.2.2.
3/1/70	4/7/70	$LE = -82.0 + 61.43 \sin(\theta + 320^{\circ})$	6.2.3.
16/1/70	16/7/70	$P = -58.1 + 62.67 \sin(\theta + 333^{\circ})$	6.2.4.

TABLE 6.4

RESULTS OF HARMONIC ANALYSIS OF ORIGINAL SERIES WITH THE FIRST HARMONIC EXTRACTED

Harmonic Number	No. of Days in One Cycle	% Variance Explained: Rn	A	LE	P
1	365.00	1.55	0.32	0.58	0.81
2	182.50	0.45	41.87	2.83	3.06
3	91.25	0.72	6.76	2.50	1.03
4	45.63	3.35	6.70	4.11	5.61
5	22.82	0.02	6.08	0.53	0.99
6	11.41	1.43	1.35	4.04	3.92
7	5.71	1.79	0.97	1.04	2.01
8	2.86	0.78	0.09	0.76	4.00
9	1.43	0.05	0.22	1.65	1.42
10	0.72	0.67	0.46	1.71	0.47
Total Explanation by first 10 harmonics		10.81	64.82	19.75	23.32

to spectral analysis using the method outlined by Blackman and Tukey (1959 p.52-54). The functions given by these authors were used to (1) prewhiten the series, (2) calculate the autocovariance, (3) perform the finite cosine series transform, (4) smooth the raw spectral estimates, and (5) correct for prewhitening. The use of 365 data points restricts the analysis to cycles of between 182.5 and 2 days in length. In the actual analysis lags from 1 to 60 days were used, and the powers of cycles ranging from 120 to 2 days in length were computed.

Confidence limits cannot be placed on the analysed power spectra as, strictly, the population is unknown. However, this is not uncommon in spectral analyses of this kind of meteorological data (see for example Dickson, 1971), and peaks or troughs, or more especially groups of relatively high or low powers, can indicate the likelihood or unlikelihood of variations with certain average periods (Panofsky and Brier, 1958 p.146).

Interpretation of the present spectra was made in the light of the above comments. However, as a further interpretive aid, a spectral analysis was made of a synthetic time series of values similar to those of the heat balance components, but containing known periodicities. The function used to create the series was

$$X_t = M + A_1 \sin \left(\frac{2\pi t}{365} + \frac{2\pi t_i}{365} \right) + A_2 \sin \left(\frac{2\pi t}{w_1} + \frac{2\pi w_1}{365} \right) + \ldots \qquad 2.2.5.$$

In this equation X_t is the value of the variable X at day t.

M is the mean value of the series. A_1 is the amplitude of the first harmonic of the series, while $A_{2,3\ldots}$ etc. are the amplitudes of the higher frequency cycles of periods w_1, $w_2\ldots$ days. t_i and w_i also define the phase angles of the first cycle and higher frequencies. Cycles of 365, 6, 5, 4 and 3 days were used. Values of A_1 were taken from the amplitudes of the first harmonic of the respective heat balance components series, and the amplitudes of the higher frequencies were all 0.33 of A_1. The value of the phase angle of the first harmonic was obtained by inserting, in t_i, the value (in days) from the phase angles given by equations 6.2.1. - 6.2.4. Similarly, the values of M were obtained from these equations. The values of the power of the synthetic series are shown on the spectra of the real data in Figs. 6.1 - 6.4, for which Table 6.5 shows the period length in days associated with each lag. In the spectra, the log of power was plotted in order to clarify details on the higher frequency cycles.

Once more a full discussion of the results is delayed until section 6.2.d. In brief, the results from all four spectra showed low frequencies to be clearly marked. These were presumably partially derived from the influence of the annual cycle. The soil heat flow spectrum was different from the other three in that it had only one clearly marked cycle at higher frequencies. In the other spectra a variety of high frequency cycles appeared to be important, with a cycle of length 4.138 days being exhibited in all of them. In all spectra powers of frequencies higher than 2.727 days had larger values than those of the respective synthetic series.

TABLE 6.5

LENGTH OF PERIOD IN DAYS FOR EACH LAG NUMBER IN POWER SPECTRA

Lag	Period (Days)	Lag	Period (Days)
0	0.0	30	4.000
1	120.000	31	3.871
2	60.000	32	3.750
3	40.000	33	3.636
4	30.000	34	3.529
5	24.000	35	3.429
6	20.000	36	3.333
7	17.143	37	3.243
8	15.000	38	3.158
9	13.333	39	3.077
10	12.000	40	3.000
11	10.909	41	2.927
12	10.000	42	2.857
13	9.231	43	2.791
14	8.571	44	2.727
15	8.000	45	2.667
16	7.500	46	2.609
17	7.059	47	2.553
18	6.667	48	2.500
19	6.316	49	2.449
20	6.000	50	2.400
21	5.714	51	2.353
22	5.455	52	2.308
23	5.217	53	2.264
24	5.000	54	2.222
25	4.800	55	2.182
26	4.615	56	2.143
27	4.444	57	2.105
28	4.286	58	2.069
29	4.138	59	2.034

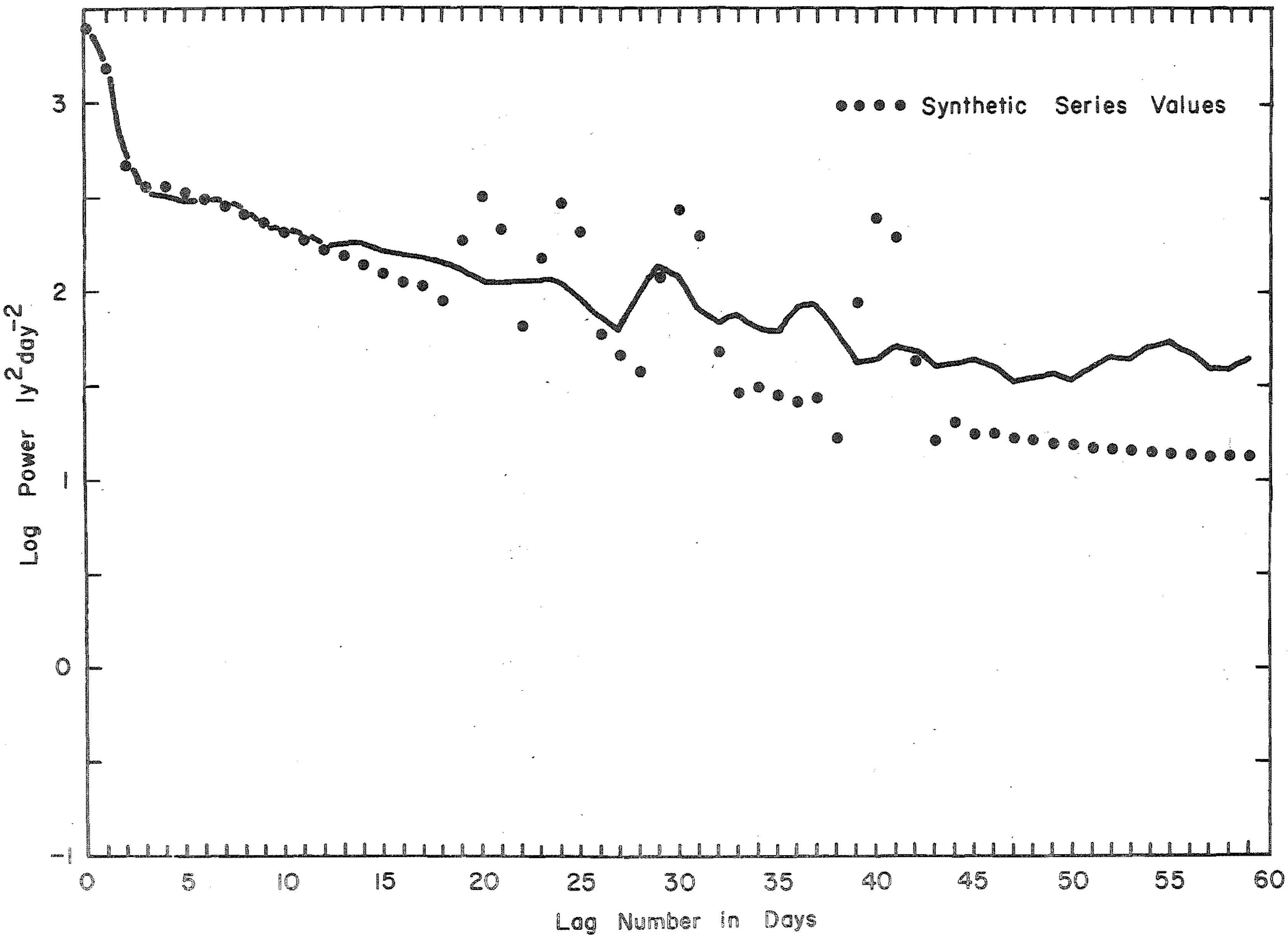


FIGURE 6.1: Power spectrum of daily values of Net Radiation

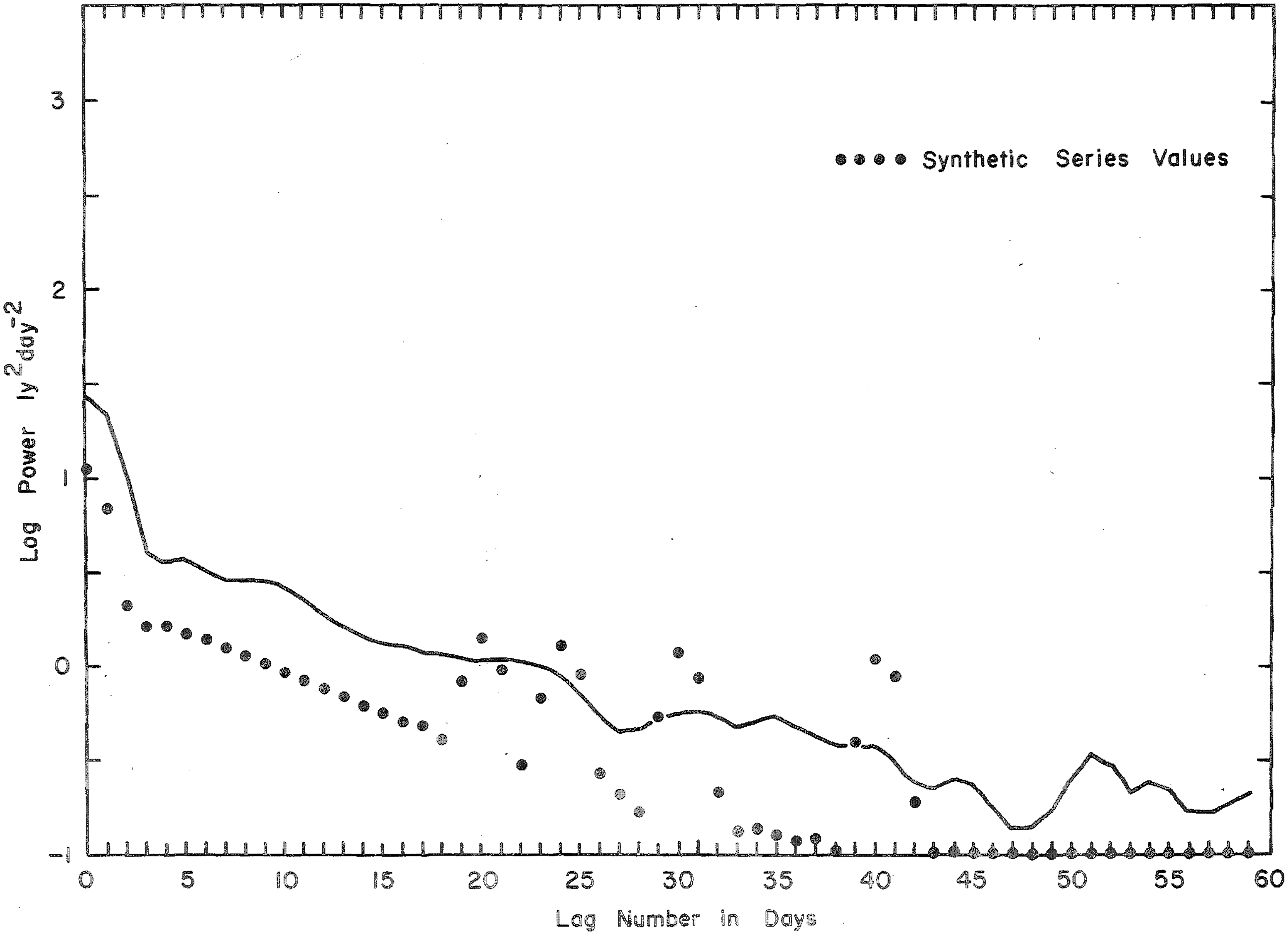


FIGURE 6.2: Power spectrum of daily values of Soil Heat Flow

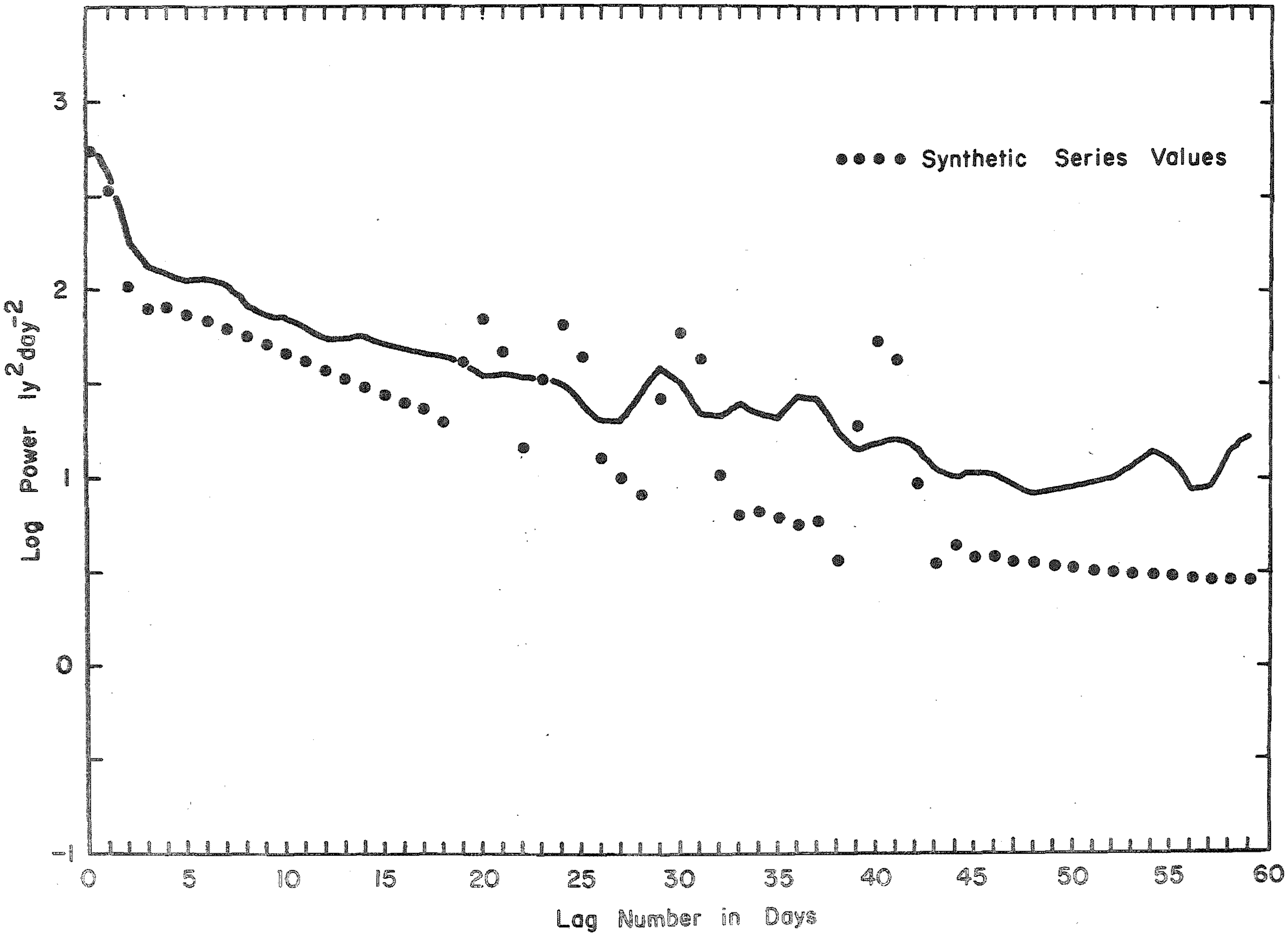


FIGURE 6.3: Power spectrum of daily values of Evaporative Heat Loss

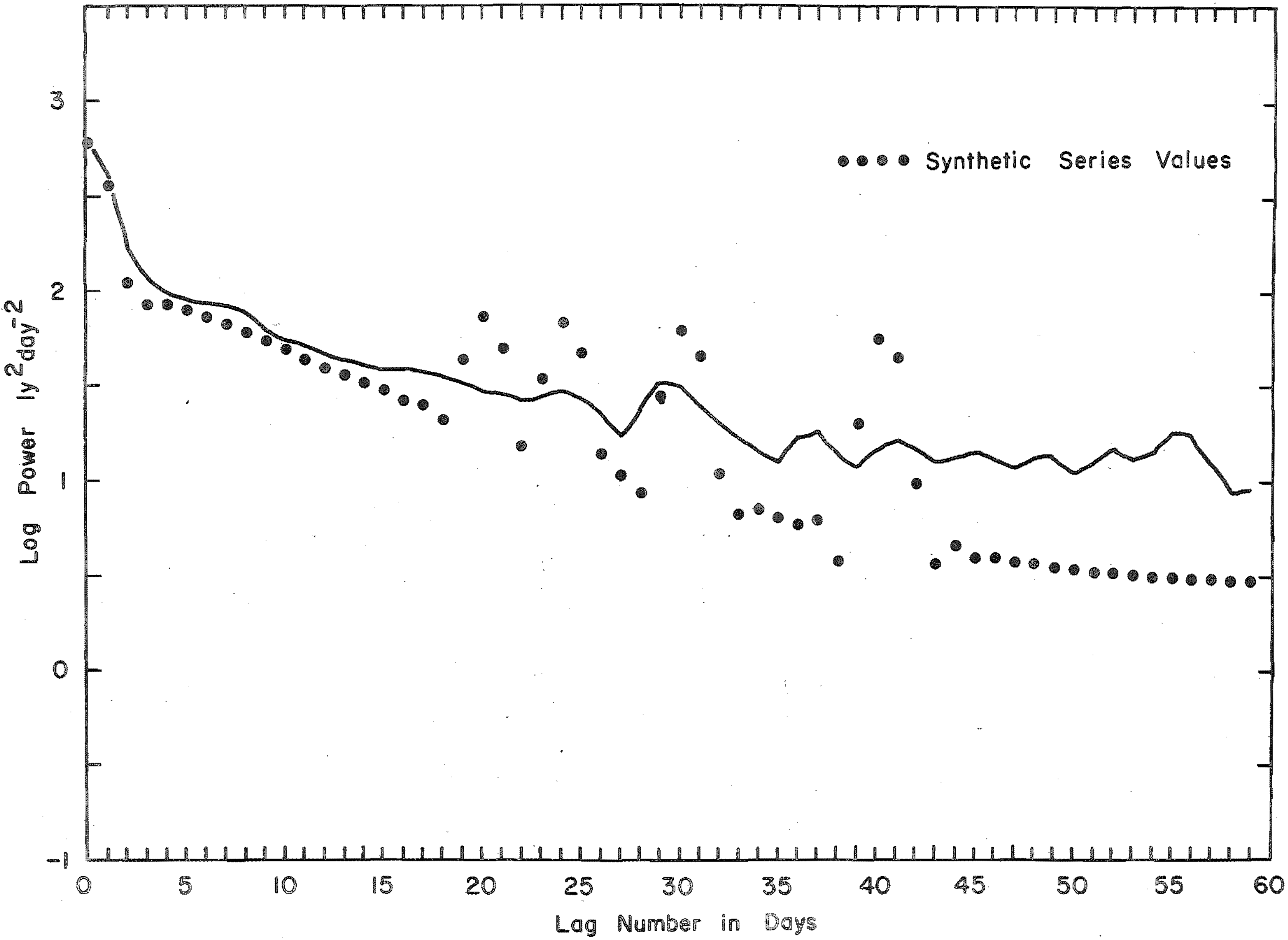


FIGURE 6.4: Power spectrum of daily values of Sensible Heat Flow

As in the harmonic analysis, values from the first harmonic of each series were subtracted from the original series in order to investigate more closely the higher frequencies. The main results from the spectral analysis of the new series are shown in Table 6.6, where the powers of apparently important frequencies are listed. In the spectra of all the series, low frequencies were still the most important, but in all except the soil heat flow spectrum, peak power was centred on a cycle of 60 days rather than 120 days. Also in all except the soil heat flow spectrum a cycle of 4.1 days was again exhibited. Some evidence of higher frequencies being present was shown in the spectra of Rn, LE and P, and in the latter a cycle of 17.1 days appeared. The soil heat flow spectrum showed only one peak at low frequency and almost 'pure noise' at all other frequencies examined.

6.2.c. Spectral Analysis of Other Parameters

In the above analyses, owing to the use of daily data, it was impossible to obtain information on cycles of less than two days in length. The analyses could not be extended in terms of the heat balance components, as data at intervals more frequent than a day were unavailable. However, hourly values of air temperature and 7½ minute records of SW↓ were available, and part of these data were analysed in order to investigate the importance of the daily cycle, and to determine if any other shorter cycles were important.

Spectra of temperature values were computed for the months of December 1969, and March and June 1970. The spectra of December and March showed only three marked peaks. Table 6.7 shows that these were centred around cycles of 10, 1 and ½ days

TABLE 6.6

LOG POWERS OF OUTSTANDING FREQUENCIES IN SPECTRA OF DAILY VALUES OF HEAT BALANCE COMPONENTS AFTER VALUES OF THE FIRST HARMONIC OF THE SERIES HAVE BEEN SUBTRACTED POWERS ARE IN LY^2 DAY^{-1} PERIODS IN DAYS

Rn		A		LE		P	
Period	Power	Period	Power	Period	Power	Period	Power
60.0	221.5	120.0	16.5	60.0	115.4	60.0	103.9
17.1	171.0			4.1	36.9	17.1	53.2
4.1	127.5			2.0	17.8	4.1	31.4
3.2	84.2					2.2	19.1
2.2	59.6						

TABLE 6.7

LOG POWERS OF IMPORTANT FREQUENCIES IN HOURLY AIR TEMPERATURE VALUES. POWER IN $^{o}C^{2}$, PERIOD IN HOURS

December		March		June	
Period	Power	Period	Power	Period	Power
240.0	1.503	240.0	2.114	240.0	2.743
24.0	1.895	24.0	1.098	24.0	0.411
12.0	0.214	12.0	0.256		

TABLE 6.8

LOG POWERS OF IMPORTANT FREQUENCIES IN 7½ MINUTE INCOMING SHORT WAVE RADIATION VALUES. POWERS IN LY^2MIN^{-2}. PERIOD IN HOURS

December		March		June	
Period	Power	Period	Power	Period	Power
24	-0.142	24	-0.501	24	-1.882
12	-0.899	12	-1.091	12	-2.153
8	-1.669	8	-1.905	8	-2.574
				6	-2.950

in length. The two former cycles were also seen in the June temperature spectra.

Spectra of SW1 were computed for the same months as the temperature analyses. In all three months (Table 6.8) the low frequency cycles of 24 hr in length were the most important, with sub harmonics of 12 hr, 8 hr and, in the June case, 6 hr, also being well marked. In all three spectra the higher frequencies showed peaks of relatively small power centred on periods o 36-45 minutes, and on periods of 21 or less minutes. Both groups were apparent in the December and March spectra but only the first was seen in the June spectrum.

6.2.d. <u>Discussion of Results of Time Series Analyses</u>

Although quantitative methods of time series analysis have been used, the interpretation is of necessity at least partly qualitative. It is for this reason that the discussion of the results has been left until this section, in order that the broad picture is obtained rather than concentrating on individual results. The more important cycles that have emerged from the harmonic and spectral analyses of the values of the heat balance components are summarised in Table 6.9.

The initial harmonic analyses of the heat balance components showed the annual cycle to be important in all the series. This was especially so in the Rn series, but in the A series the second harmonic was more marked. Higher frequencies with cycles of less or equal to 11.4 days were seen only in the LE and P series. When the values from the first harmonic were removed from the series, the lower frequencies

TABLE 6.9

MAIN CYCLES IN THE HEAT BALANCE COMPONENT TIME SERIES SELECTED BY THE DIFFERENT ANALYSES. CYCLES ARE IN DAYS

	Rn	A	LE	P
Harmonic Analysis	365.0	365.0	365.0	365.0
	45.6	182.5	182.5	182.5
		91.3	91.3	45.6
		45.6	45.6	11.4
		22.8	11.4	2.9
Harmonic Analysis with 1st Harmonic subtracted	365.0	182.5	18.25	182.5
	45.6	91.3	91.3	91.3
	11.4	45.6	45.6	45.6
	5.7	22.8	11.4	11.4
		11.4	5.7	5.7
			1.4	2.9
			0.7	1.4
Spectral Analysis	120.0	120.0	120.0	120.0
	4.1	2.4	4.1	4.1
	3.2		2.0	2.2
Spectral Analysis with 1st Harmonic subtracted	60.0	120.0	60.0	60.0
	17.1		4.1	17.1
	4.1		2.0	4.1
	3.2			2.2
	2.2			

still remained important in all the series, especially that of A, and a cycle of 45.6 days was noteworthy in each series. More emphasis was given to the higher frequencies in the LE and P series with the removal of the first harmonic values. The well marked annual cycle shown by the importance of the first harmonic of the Rn, LE and P series, and the fact that the harmonics were in phase within 13 days, may both be considered characteristics of the dynamic climatology of this location for the period studied.

The spectral analyses complement the harmonic analyses. Once again low frequencies were dominant, although in this case the lowest frequency able to be selected is one of 120 days. However more weight may be attached to the ability of this method to select possibly important high frequency cycles, and in this respect it is noteworthy that a cycle centred around 4.1 days appeared to be important in the Rn, LE and P series. Other high frequency cycles appeared to occur in these series but none was as important as the above. It is interesting to note, however, that the power values at high frequencies in all the series were relatively large, and this is, to some extent, a measure of the day to day variability in the daily values of the heat balance components that has been alluded to in earlier chapters. The spectra of the Rn, LE and P series, with the first harmonic values removed, pointed to a 60 day cycle being more important than the 120 day cycle. Together with the results of the harmonic analysis, this suggests that any low frequency cycles, other than the annual one, that existed in the series had periods ranging between about 46 to 91 days. In most of the analyses the soil heat flow series showed different characteristics from the other series. Reasons for this may include (1) the small absolute values of the data, (2) the fact that daily variation may be of the same order as the seasonal variation, and (3) the slower rate of heat flow through the submedium compared with that through the air.

Meteorological time series frequently show continuous spectra with low frequencies being important and the annual and diurnal cycles being statistically significant (Ward and Shapiro, 1961). Although confidence limits have not been added, the present results, including the diurnal cycles as discussed below, are in general agreement with this conclusion. Assuming, for the sake of comparison, stationarity of meteorological time series, and the relation of parameters such as air temperature and pressure to the values of the heat balance components, the present results may be compared with those of Rayner (1965). Rayner analysed many series of meteorological variables from different locations in New Zealand for the year 13 August 1962 to 12 August 1963. Besides the diurnal and annual cycles he found evidence for the existence of cycles of 40, 16, 6.7 and 5.0 days. None of these latter cycles is directly comparable to those found in the present study although the following similarities in minor details between the results of the two works may be noted.

(1) Frequencies corresponding to 40 day cycles always have high powers in the heat balance analyses.

(2) Low frequency cycles are in general well marked in the heat balance spectra.

(3) In an analysis of SW↓ for Christchurch Rayner found marked peaks at 3.3 and 2.8 days, the former being comparable to the cycle of 3.2 days found in the Chilton Valley Rn spectra.

(4) In the Rn, LE and P spectra there is a noticeable decline of power after a cycle of period between 17.1 and 15.0 days, which may bear some relation to Rayner's 16.0 day cycle.

However, there is little evidence in the heat balance spectra of cycles comparable to those of 6.7 and 5.0 days, which Rayner tentatively related to the passage, at regular intervals, of synoptic cyclones and anticyclones.

With regard to the comparison of the results of the two studies, two features of recent work by Dickson (1971) are relevant. Firstly, Dickson was able to relate the spectra of daily mean surface air temperature, in the United States, with seasonal mean 700 mb heights in a grid extending from mid-Atlantic to mid-Pacific. He concluded that power was concentrated in progressively longer period temperature oscillations as the seasonal mean circulation proceeds from fast westerlies, in the vicinity of any one station, to highly amplified wave patterns. Assuming that there is a relation between mean daily surface air temperature and daily values of heat balance components, the relative dominance of low frequencies in the present data, and the absence of the 6.7 and 5.0 day cycles of Rayner, in the light of Dickson's work, suggest that the present data may be associated with large amplitude wave patterns in the upper air. Evidence that these in fact existed, at least in the summer months of the study period, has been given by Hill (1971). Secondly, Dickson also evoked the concept of Landsberg et.al. (1959) that most of the seasons have their own characteristic spectral signature, and that the spectra for a given season vary widely from one year to another. Extension of this to an annual basis would suggest that the year studied by Rayner and the year of the present data demonstrated their own characteristic spectral signatures. On the basis of Dickson's work, the explanation of the variation might be found in the differences of the mean upper air circulations of the two years.

If the different periodicities, 6.7 and 5.0 days on the one hand, and 4.1 days on the other, are related to the passage of anticyclones across New Zealand, then the commonly quoted peridocity for this passage of 'about a week' (de Lisle, 1969) may be placed in perspective in two ways. Firstly, quantitative data on the extent of the generality of the periodicity has become available. Secondly, since Rayner found that the statistically significant 6.7 and 5.0 day cycles only explained about 20% of the total variance, it is recognised that the real situation cannot be so readily generalised. However, a more specific study of the frequency of the passage of anticyclones would have to be made to confirm these points.

Attention is now turned to the spectra of air temperature and SW↓ values. An outstanding feature in all the three months analysed was the daily cycle appearing in both series of parameters. Although it was the most important in the radiation spectra, it took second place to a ten day cycle in the three temperature spectra. The ten day cycle and the principal higher frequency cycles, found both in the temperature and radiation spectra, are interpreted here not as representing the existence of physically real cycles of these periods, but rather, as in the case of atmospheric tides (Butler and Small, 1963), representing the interaction of sub harmonics of the main diurnal periodicity. It is possible that the very high frequency cycles (less than 45 minutes) in the SW↓ spectra may have some physical cause. For example the distribution of skyline ridges and troughs in the valley might influence the values of SW↓ in a regular fashion. But again further study would be necessary to verify this hypothesis.

Overall, as a result of the above time series analyses two general conclusions may be drawn. Firstly, astronomically controlled cycles, together with other cycles that may be associated with conditions of upper air flow, were reflected in the flows of energy at the earth-atmosphere interface at the present location. Secondly, the existence of such cycles is an integral part of the climate of the Chilton Valley and emphasises the dynamic character of the climate itself.

6.3 The Climatonomy of the Chilton Valley

6.3.a. The Estimation of the Mean Annual Energy Budget

The underlying assumption of climatonomic theory is that values of air temperatures and energy fluxes, at a location, may be determined using basic physical laws and limited initial data. In the part of climatonomic theory concerned with the estimation of the mean annual energy budget three procedures are followed. These are:-

(1) The collection of basic data and the making of necessary assumptions.

(2) The derivation of the energy budget.

(3) The checking of the values of the energy budget by means of additional computations and data.

In the following analysis, two almost independent data sets are used. The first has data and assumptions that include no information collected during the study period. The second uses mainly data gathered from the study period. A comparison of the results of applying the above procedures to these data sets is then made.

The basic data is shown in Table 6.10. A few notes about the data external to the study period are necessary. The Cass precipitation refers to data collected from the Biological Station. There are few New Zealand data on albedos and Bowen ratios, therefore the albedo value adopted is one assumed by Jackson (1967) for North Island grassland. The assumed Bowen ratio is computed from the same source, assuming values of Rn and PET for a 0° slope, and also assuming zero net soil heat flow for the year. No runoff-precipitation ratio can be computed for the Waimakariri Catchment, in which the Chilton Valley is located, as river discharge cannot be accurately assessed. The Selwyn Catchment, south of the Waimakariri, is much smaller (152 km^2) and is composed principally of alpine foothills. The quoted ratio is the average for the years 1962-68, and is computed from data published by the Ministry of Works (1962-66) and supplied by G. Martin (pers. comm.) for the years 1967-68.

The steps given by Lettau (1968) for computing the mean

TABLE 6.10

BASIC DATA FOR THE ESTIMATION OF THE MEAN ANNUAL HEAT BUDGET OF THE CHILTON VALLEY

Variable	Source	Value
Incoming SW Radiation	Estimated from de Lisle (1966)	330 ly day^{-1}
	Recorded Chilton Valley	276 ly day^{-1}
	Chilton Valley adjusted for sky line obstruction	322 ly day^{-1}
Monthly Precipitation	Cass (Greenland and Owens, 1967)	10.9 cm $month^{-1}$
	Chilton Valley study year	8.3 cm $month^{-1}$
Runoff/precipitation ratio	Selwyn Catchment (NZ MOW 1962-8)	0.58
	Chilton Valley (Soons, 1970)	0.02
Latent Heat of Vapourisation		595 cal gm^{-1}
Albedo	Taita (Jackson 1967)	0.25
	Chilton Valley measured	0.13
Bowen Ratio	Taita, computed from Jackson (1967)	0.20
	Chilton Valley	0.61

annual energy budget are as follows. Net SW↓ is computed from SW↓ $(1 - \alpha)$. LE is taken as the amount of heat needed to evaporate the annual rainfall multiplied by (1 - runoff/ precipitation). P is computed from LE multiplied by the Bowen ratio. Net LW is given by the sum of net SW, P and LE, where the latter two already have negative signs. Net LW is given a negative sign to conform with the convention being used in this study. Rn is then the sum of net SW and net LW. The results of the application of these steps (Table 6.11) will be discussed after the checking computations have been described.

A check on the heat budget estimates can be made by means of computing the mean air temperature from the value of net LW. The data for this calculation, the assumptions used, and the method of computation are given in Appendix D. The results in the form of mean annual air temperature at 2 m are shown in Table 6.11. The check can only be used as a guide, since air temperature itself is used in the computation. Moreover, the data are drawn from different locations, and relate to different years. However, the results indicate that heat budget estimates 1 and 4 are in the right order of magnitude, while estimates 2 and 3 are in error.

In the discussion of the results (Table 6.11) of the application of this part of climatonomic theory, it should be pointed out that estimates 1, 2 and 3 apply to the study year, whereas estimate 4 generally uses data averaged over a period of years, and therefore gives an estimate for a 'normal' year. In general, estimate 4 compares quite favourably with estimate 1, indicating that

TABLE 6.11

ESTIMATE OF ANNUAL HEAT BUDGET OF CHILTON VALLEY USING DIFFERENT ASSUMPTIONS AND DATA SOURCES. VALUES IN LY DAY^{-1}

		Net SW Rad.	A	P	LE	Net LW Rad	Rn	$\bar{T}$ 200 °C
1	Estimates from previous chapters	244	2	-58	-82	-108	136	-0.8
2	Estimates using recorded Chilton Valley data in Table 6.10	240	0	-89	-146	- 5	235	-147.1
3	Estimates using Chilton Valley data in Table 6.10 with SW radiation adjusted for sky line obstruction	281	0	-89	-146	-46	235	-53.6
4	Estimates using data from other sources	248	0	-21	-103	-124	124	9.2

this section of climatonomic theory can be applied successfully to this location. The major differences between estimates 1 and 4 are in the value of LE, and the partition of heat loss between P and LE. The difference in the value of LE is due to the fact that the study year had below average rainfall. The difference in the partition of heat between P and LE simply reflects the different values of Bowen ratio that were applied. Estimates 2 and 3, from the Chilton Valley during the study period, are in error due to the low runoff-precipitation ratio that was applied. This indicates that for the successful application of this part of climatonomic theory, the runoff-precipitation ratio chosen should not be that referring to small plots, but should be that for the whole catchment area. The latter includes water which was lost to evaporation owing to deep percolation, but by means of ground water flow, may find its way into rivers and form part of the catchment discharge value. In order to obtain a reasonable value of LE, using climatonomy, in estimates 2 and 3, a runoff-precipitation ratio of 0.56 would have to be used, and this is very similar to the value quoted for the Selwyn catchment.

6.3.b. Estimation of the Annual Variation of the Heat Budget

A summary of the basic theory of Lettau's (1968) method of estimating the annual variation of the heat budget by means of climatonomy is given in Appendix C, and further details are presented in Appendix E. The necessary steps for the application of the theory are as follows. Firstly, the forcing function, net SW, is expanded as a Fourier series.

Then the average annual surface temperature is calculated using response parameters. This is followed by the calculation of the response parameters needed to compute the annual variations. The latter parameters include those that are independent of season, and those which vary with the different harmonics of the annual cycle. The variable parameters include the important sets of partial impedances, and phase constants, of the heat balance components. Finally, a Fourier synthesis of the response functions is made to give the annual variations.

For the purposes of this analysis the original monthly mean values of net SW for the study period (Table 6.1) were reduced to eight equally spaced coordinates (Table 6.13). The reduction was made in the following way. Where the data of the coordinate fell on the first of the month, the mean of the mean monthly value of the two adjacent months was taken. Where the coordinate fell in mid month, the mean value for that month was adopted. As an approximation, the analysed value for August was taken as the average value of the means of the parts of August 1969 and 1970. The analysis was performed using principally data derived from the Chilton Valley during the study year. Therefore the values of the heat balance estimated by climatonomic theory should be comparable to the measured or estimated heat balance values for the study year. The analysis was repeated using one major variation. All other factors being the same (except for a slight change in A), the lower value for the Bowen ratio (0.20), which was employed in section 6.3.a., was adopted. This serves the purposes of providing information on the possible variation of the values of the heat budget, as will be seen in the later discussion (section 6.3.c.).

The Fourier expansion of the forcing function, F, for the study year is

$$F = 2.95 + 2.36 \cos (nt + 223) + 0.098 \cos (2nt + 169) + 0.118 \cos (3nt + 177) + 0.285 \cos (4nt + 180) \qquad 6.3.1.$$

where F is in m ly sec^{-1}, n equals 20.10^{-8} sec^{-1}, and t equals zero at August 15, 1969. As a simplifying assumption, the mean annual surface temperature, in the present analysis, is taken as $9.4^{o}C$, which is computed directly from the measured mean air temperature of $8.8^{o}C$ using the method described in Appendix D. The reasons for making this assumption are that the present analysis is concerned with annual variation rather than the annual means, and that errors in calculations $\bar{\bar{T}}_o$, due to incorrect assumptions, such as that of the value of the Angström ratio, will not be present in $\bar{\bar{T}}_o$. The values of the response parameters used in the computation of annual variations, and which are assumed not to vary with season, are as follows. Caloric admittance of the submedium, μ, is 14 m ly deg^{-1} sec^{-1}; $c_p \bar{\rho}$ is 0.27 m ly deg^{-1} cm^{-1}, u_* is 27.2 cm sec^{-1}, and the phase angle of A, ψ is 45^{o}. All relevant derivations, parameterisations, and assumptions used are shown in Appendix E.

The values of the response parameters that vary with season and are different in the individual harmonics of the annual cycle are shown in Table 6.12. The corresponding values when the Bowen ratio is 0.20 are given in Appendix E, where again derivations are presented. The Fourier synthesis to calculate

TABLE 6.12

VALUES OF PARAMETERS THAT VARY WITH THE FIRST TO FOURTH HARMONICS. IMPEDANCES, AND AMPLITUDES OF HEAT BALANCE COMPONENTS ARE IN M LY SEC^{-1}. PHASE ANGLES ARE IN DEGREES

Harmonic	$\Gamma + \beta$	ψ	N	Φ
1	0.02088	0.00630	7.1793	0.09511
2	0.02184	0.00896	6.8781	0.10560
3	0.02279	0.01092	6.7022	0.11250
4	0.02375	0.01260	6.5771	0.11820

Harmonic	φ	ζ	z	ΔT_O
1	6.8	7.1	0.2788	8.465
2	7.6	8.1	0.3075	0.319
3	8.1	8.7	0.3283	0.359
4	8.5	9.2	0.3457	0.824

Harmonic	$\Delta(LW\uparrow + LW\downarrow)$	ΔA	ΔP	ΔLE
1	0.1767	0.053	0.805	1.320
2	0.0070	0.003	0.034	0.056
3	0.0081	0.004	0.040	0.066
4	0.0195	0.010	0.097	0.159

the annual variation of T_o, (LW↑ + LW↓), A, P, LE and T_{200}, the air temperature at 2 m, was made using respectively equations C.5, E.26, E.14, E.15, and E.16. The computation of T_{200} from T_o was made using the method described in Appendix D. Climatonomically estimated values of the heat budget are shown in Table 6.14, and the annual variations when the Bowen ratio is 0.2 are given in Table 6.15. In the latter table values were estimated assuming the net annual A value to be zero.

6.3.c. Discussion of the Estimates of the Annual Variation of the Heat Budget

In the discussion of the comparison of the results of the climatonomic estimates (Tables 6.14 and 6.15) with the original data (Table 6.13), the analysis with the Bowen ratio of 0.20 will be termed the experiment, and the first analysis will be referred to as the prediction. All of the series of both predicted and experimental results show generally reasonable values when compared with the original series. The main differences lie in the amplitudes of the series.

The original net LW series shows a difference of 137 ly day^{-1} between maximum and minimum values, whereas the predicted results show a range of 29 ly day^{-1}, and the experimental values have a range of only 14 ly day^{-1}. Relatively low ranges are again apparent in the predicted and experimental values of soil heat flow. The former are closer to the original records, but this is partly owing to the choice of zero net soil heat flow in the experimental analysis. Neither the predicted nor the experimental values show the large soil heat flow to the surface indicated in the original series. However, the

TABLE 6.13

REDUCTION OF ORIGINAL HEAT BALANCE DATA. HEAT BALANCE AND RADIATION VALUES ARE IN LY DAY^{-1}. FLOW TOWARDS SURFACE IS POSITIVE

	15 Aug	1 Oct	15 Nov	1 Jan	14 Feb	1 Apr	15 May	1 July
Net LW	-73	-121	-180	-173	-156	-96	-59	-43
A	+13	+11	-13	-11	+2	+5	+3	+3
P	-25	-68	-153	-103	-87	-34	-22	-2
E	-53	-97	-102	-143	-140	-76	-29	-17
T_{200} °C	4.8	6.6	12.8	14.3	13.1	11.1	3.7	3.8
Net SW	142	279	449	430	380	209	104	58
Rn	69	158	269	257	224	113	45	15

TABLE 6.14

CLIMATONOMIC ESTIMATES OF HEAT BALANCE DATA, FOR THE STUDY YEAR. HEAT BALANCE AND RADIATION VALUES IN LY DAY^{-1}. FLOW TOWARDS SURFACE IS POSITIVE

	15 Aug	1 Oct	15 Nov	1 Jan	14 Feb	1 Apr	15 May	1 July
Net LW	-95	-110	-118	-124	-115	-109	-96	-96
A	+3	-2	-2	-2	+3	+5	+7	+6
P	-1	-77	-108	-133	-92	-54	+3	-1
E	+11	-114	-164	-206	-138	-76	+18	+12
T_o °C	0.8	9.1	13.9	17.2	12.6	8.7	1.5	0.9
T_{200} °C	0.8	8.2	12.7	15.7	11.5	8.1	1.5	0.9
Net SW	142	279	449	430	380	209	104	58
Rn	47	169	331	306	265	100	8	-38

TABLE 6.15

CLIMATONOMIC ESTIMATES OF HEAT BALANCE DATA FOR STUDY YEAR WITH BOWEN RATIO OF 0.20 AND MEAN ANNUAL A OF ZERO. HEAT BALANCE AND RADIATION VALUES IN LY DAY^{-1}

	15 Aug	1 Oct	15 Nov	1 Jan	14 Feb	1 Apr	15 May	1 July
Net LW	-118	-125	-129	-132	-128	-124	-118	-118
A	+1	-2	-2	-2	0	+1	+2	+1
P	+5	-30	-44	-57	-37	-19	+7	+6
E	+30	-149	-220	-278	-183	-93	+40	+31
T_o °C	4.8	8.8	10.7	12.4	10.8	8.4	5.1	5.0
T_{200} °C	4.9	8.5	10.2	11.7	10.2	8.2	5.2	5.1
Net SW	142	279	449	430	380	209	104	58
Rn	24	154	320	298	252	85	-14	-60

climatonomic estimates do exhibit relatively low absolute values of A, rather than the higher ones suggested by the soil temperature records (section 3.4).

The effect of the different Bowen ratio values is seen clearly in the predicted and experimental series of P and LE. While the predicted values are generally of the same order as the original values, the experimental series of P and LE values are respectively lower and higher than the original values. Neither the predicted nor the experimental series appear to show evidence of the results of lack of soil moisture during November. Both series exhibit a maximum value at January 1, which is presumably due to the influence of the first harmonic. Both the predicted and experimental P and LE series show larger amplitudes than the original series, and small flows (except for the experimental LE values) to the surface in winter. The positive winter flows of P are physically reasonable, but those of LE occur owing to the absence of adjustment to the inverse Bowen ratio (see Appendix E).

Resulting values of Rn in the predicted and experimental series again show larger ranges than the original values, and demonstrate well marked outflows of Rn in winter. The range of predicted air temperatures is higher than that of the original series, while that of the experimental series is lower. The predicted air temperatures are closer to the measured air temperatures from October 1 to February 14, but the experimental air temperatures are nearer the original values at the remaining coordinates which refer to the colder half of the year.

Three general points may be drawn from the above description of the results of climatonomy. Firstly, the climatonomic model is fairly successful in giving a general pattern of the values of the heat balance components, and related factors, throughout the year at this location. Since the predicted values are generally more realistic when compared with the original values, the degree of success of climatonomy, as might be expected, depends on the validity of the initial data and assumptions, at least insofar as the Bowen ratio value is concerned. The forcing function is the most important parameter in using climatonomic theory, and the relative success in the present location suggests that the theory could be useful in making estimates of the heat balance in other parts of the High Country, if values of F were available.

The second point concerns the generalisation of the climatonomy approach. In the present analysis, where only eight coordinates and four harmonics are used, the model fails to predict values, for any one coordinate, that are markedly different from those given by the first harmonic. The values for November are exemplary in the present case. This difficulty might be overcome by using a larger number of coordinates and harmonics. However, on the basis of the above analyses, climatonomy, in its present form, appears to be more suitable for predicting climatic normals, the purpose for which it was originally intended, rather than to predict the heat balance values for any one year.

Thirdly, the value assumed for the Bowen ratio has been shown to be very important in determining the values of the predicted series. The experimental results show that in some years a Bowen ratio value of 0.20 might not be unreasonable, especially if an adjusted increase Bowen ratio value was used. Indeed, if the predicted temperature values are used as a criterion, then the value of 0.20 was more appropriate for the winter months. This is understandable as the actual Bowen ratio values for most of these months (Table 5.1) were below 0.20. This suggests that, as a development of the model, a parameterisation of the Bowen ratio, so that it might vary through the year, would be advantageous. A further extension would be for F and precipitation values to act as dual forcing functions. A rather similar approach has been suggested by Lettau (1969). The use of the lower Bowen ratio value, in the present analysis, has given some indication of the possible order of magnitude of maximum LE values in a year with continually high soil moisture amounts. It may be concluded from the above discussion that a feature of the climate of the study year was, in fact, the seasonal variation of the Bowen ratio, and its high value in summer.

6.4 Summary

The main points that arise from these studies of periodicities in the heat balance values and the climatonomy of the Chilton Valley are as follows:-

1. Astronomically controlled cycles, together with other cycles that may be associated with conditions of upper air flow, were reflected in the flows of energy at the surface in the Chilton Valley.

2. The existence of these cycles is an integral part of

the climate of the Chilton Valley and emphasises the dynamic character of the climate itself.

3. Parts of the theory of climatonomy were applied with a fair degree of success, and the climatonomy of the Chilton Valley for the study year was specified by the values in Tables 6.11, 6.12, and 6.14.

4. Climatonomic estimates of the mean annual heat budget drew attention to the large loss of percolated water that might otherwise have been evaporated. Although values of percolation were previously noted (section 4.2), their importance to the heat balance was made clear by the application of the theory of climatonomy.

5. A further characteristic of the climate of the Chilton Valley in the study year, emphasised by climatonomic theory, was the seasonal variation of the Bowen ratio, and its high value in November.

6. If accurate initial data, especially of the forcing function, were available, the theory of climatonomy should prove useful in making estimates of the heat balance in other parts of the High Country.

CHAPTER SEVEN

THE HEAT BALANCE OF THE CHILTON VALLEY

7.1 Introduction

In this chapter, the interrelationships of the daily energy flows, and the effect of synoptic weather on their values, are examined. Also, a comparison with heat balance estimates for other parts of the world is made. It is shown that relationships exist between the values of the daily energy flows at the Chilton Valley, and that the strength of the relationships varies with season. The values of the daily flows are affected generally by large scale synoptic weather, although the effect is more clear in some weather situations than in others. A comparison with heat balance estimates for other parts of the world, helps to draw attention to the important features of the heat balance at the present location during the study year. As a result of all of these studies, notable features of the character of the climate of the Chilton Valley become apparent.

7.2 The Interrelationships of the Heat Balance Components

The interrelationships of the energy fluxes are demonstrated by multiple regression analysis. Analyses were performed using each heat balance component in turn as the dependent variable. Daily values for the whole study period, and for individual seasons within the study period, were analysed. The linear correlation coefficients between pairs of the heat balance components were obtained. Values of LE, P, and A were entered into the analysis with positive signs, where energy flowed away from the surface. This affects the signs of the simple correlation coefficients. The results of the analyses (Tables 7.1 - 7.5) were significant at the 95% confidence level.

In interpreting the results of the multiple regression analyses it must be remembered that certain relationships between the variables are known to be present. This is owing to the methods by which the daily values of the energy fluxes were obtained. Figure 7.1 represents diagrammatically the relations between the energy fluxes owing to the methods used to obtain their values. It can be seen that P is already related to Rn, LE and A, having been obtained as the residual of the heat balance equation (equation 1.1.1.). LE is a function of Rn although Ea, the 'drying power' term, and precipitation, also affect its value. Therefore Rn and A are the only variables that have been obtained independently of the other energy flows. P and LE are already related in some degree to Rn and A. In the light of this, fairly high correlations may be expected, at least between Rn, LE and P. Therefore attention must not only be given to the values of the linear

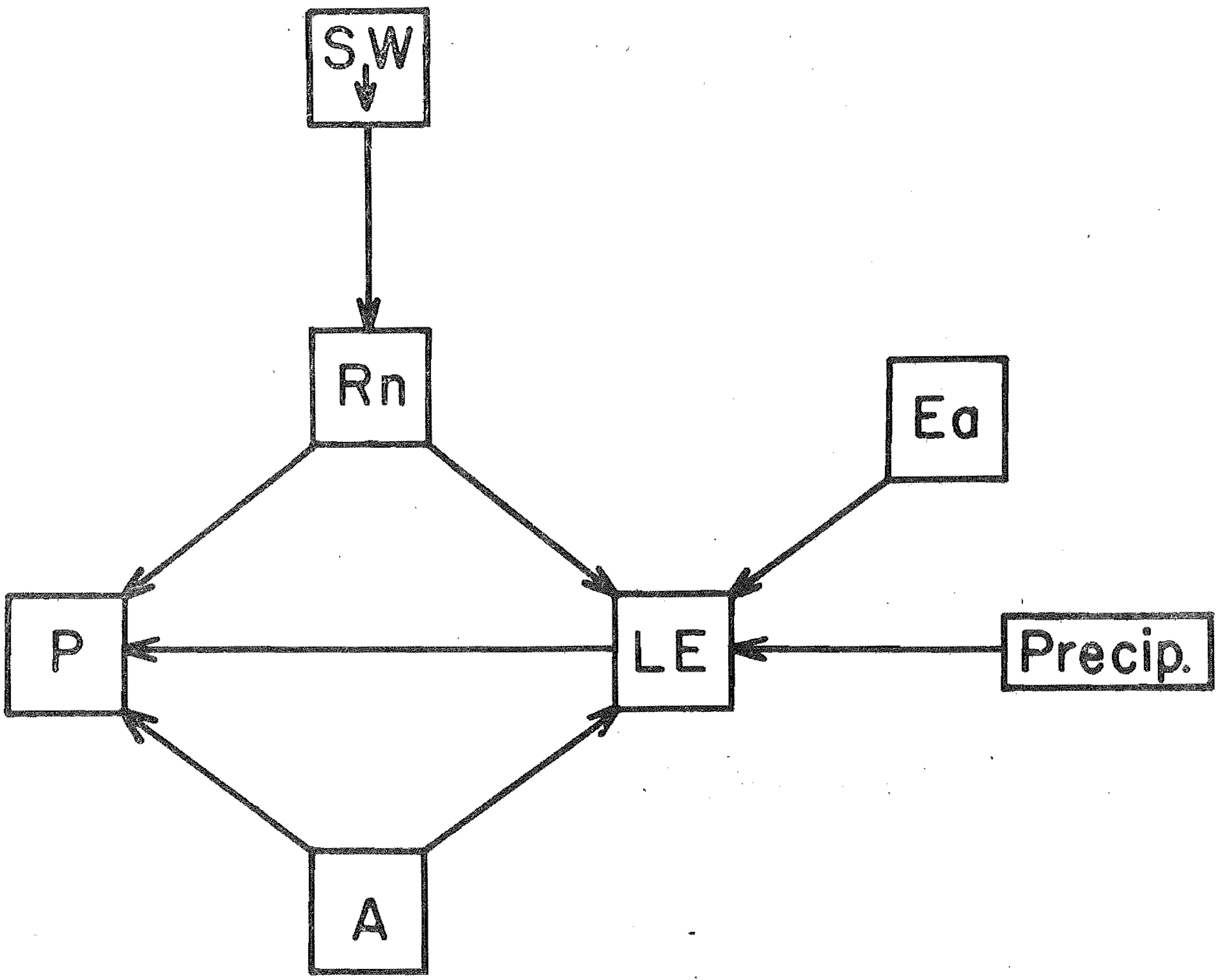


FIGURE 7.1: Diagram illustrating the relationships existing between the values of the heat balance components as a result of the methods of derivation

correlation coefficient (C.C.) and the multiple correlation coefficient (M.C.C.), but also to the sequence in which the independent variables were taken (stepwise) into the multiple regression equations. Furthermore, the departure of the multiple regression equations from equation 1.1.1. may be used as being indicative of the effect of non-radiative factors acting through LE and A. Such a departure may be seen in the values of the constant term and the individual coefficients of the multiple regression equations. Therefore, with regard to the analyses in general, although collinearity is recognised, it need not necessarily prohibit the acquisition of meaningful results.

The values of C.C. for the annual data are shown in Fig. 7.2, and those for the seasonal data will be discussed in relation to these. In the annual data P and LE appear to be most strongly related to Rn, and less strongly between themselves. The weakest relationship appears to be between A and Rn, and P and LE, and in particular the latter two. In the multiple regression equations in Tables 7.1 - 7.5, the independent variables are written in the order in which they entered into the regression. Therefore, in the annual data, Rn is seen to be most important in determining the values of A, LE and P. It is also noteworthy that A is not entered into any equations (at the 95% confidence level). Although the M.C.C. is high for most equations, that where A is the dependent variable, is very much lower. This again demonstrates the weaker relation of A to the other components.

The results of the analysis for the spring months of September, October and November, 1969 (Table 7.2), show

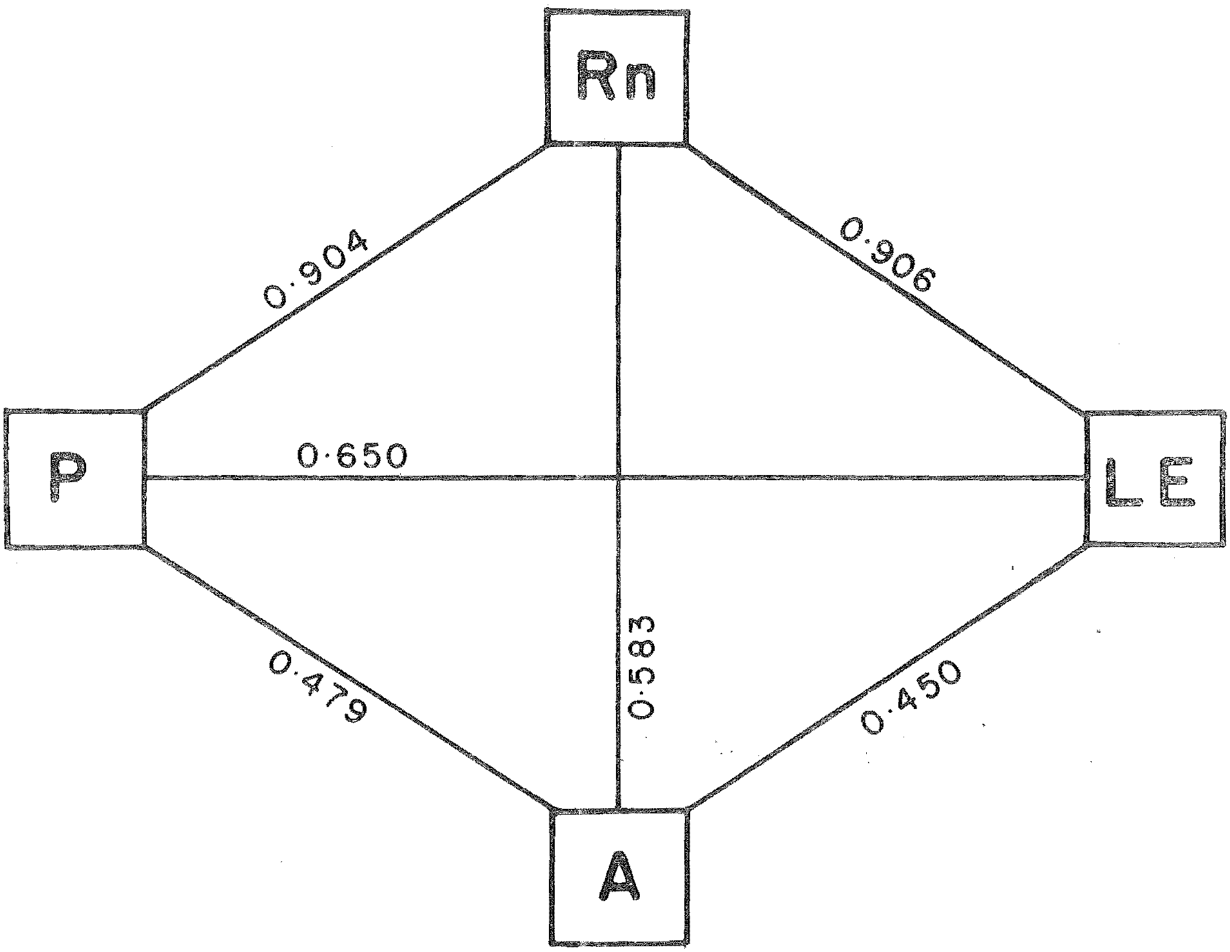


FIGURE 7.2: Simple linear correlation coefficients between the values of the heat balance components during the study year

TABLE 7.1

RESULTS OF MULTIPLE REGRESSION ANALYSIS OF ALL DAILY VALUES OF HEAT BALANCE COMPONENTS DURING THE STUDY PERIOD

Correlation Matrix

	Rn	A	LE	P
Rn	1.000	0.583	0.906	0.904
A		1.000	0.450	0.479
LE			1.000	0.650
P				1.000

Multiple Regression Equations

		Equation	M.C.C.	S.E.E.
Rn	=	-9.1389 + 1.058 LE + 1.0636 P	0.996	10.031
A	=	-8.3785 + 0.098 Rn - 0.0832 LE	0.612	9.219
LE	=	10.1428 + 0.9174 Rn - 0.9486 P	0.988	9.395
P	=	8.4927 + 0.9012 Rn - 0.9163 LE	0.988	9.233

TABLE 7.2

RESULTS OF MULTIPLE REGRESSION ANALYSIS OF DAILY VALUES OF THE HEAT BALANCE COMPONENTS FOR SEPTEMBER, OCTOBER AND NOVEMBER 1969

Correlation Matrix

	Rn	A	LE	P
Rn	1.000	0.701	0.704	0.863
A		1.000	0.397	0.555
LE			1.000	0.269
P				1.000

Multiple Regression Equations

Equation	M.C.C.	S.E.E.
Rn = -0.0246 + 1.0000 P + 0.9998 LE + 1.0032 A	1.000	0.611
A = -0.0204 + 0.9948 Rn - 0.9944 LE - 0.9946 P	0.999	0.609
LE = 0.0406 + 1.0001 Rn - 1.0001 P - 1.0031 A	1.000	0.612
P = 0.0346 + 0.9999 Rn - 0.9997 LE - 1.0029 A	1.000	0.612

generally lower simple correlation between all variables, except between A and P, and A and Rn. The lower correlations between Rn, LE and P are due to the fact that, during this period, there was a transition between the relative sizes of LE and P, since the dry month of November was associated with high values of P. The higher correlations between A and Rn may be due to the high persistence in their daily values during this period (see Appendix F). In the multiple regression equations for the data of this period, Rn again plays a dominant role in determining the values of the other three variables. Where Rn is the dependent variable P, rather than LE, is the most influential factor. This is the only season where this occurs, and is probably due to the high values of P in November. The multiple regression equations are also noteworthy for their high values of M.C.C. and low values of S.E.E., together with the fact that A is now brought into them. The equations are identical to the heat balance equation when the constant and coefficients are rounded to the nearest whole number. To summarise, the analysis of the relations between the daily values of the heat balance components for this season shows two features. Firstly, the effect of the dry period in November is seen. Secondly, since the multiple regression equations are similar to the heat balance equation, a high degree of interaction between the energy fluxes and little influence from external factors is indicated.

In the summer months of December 1969, January and February 1970, the relative magnitudes of the values of energy flows, with respect to each other, were the most stable (of all seasons). Sufficient soil moisture for most of the period, gave rise to LE having higher values than P. Values of A generally indicated flow away from the surface. They have already been shown (section 3.3) to be more closely related to Rn during November and December 1969, and January 1971, than during the whole year. This 'steady state' situation is reflected by values of C.C. (Table 7.3) which are higher in all cases (except between P and A, where the difference is small) than the corresponding coefficients from the analysis of the annual data. Once more, the multiple regression equations have high values of M.C.C. and low values of S.E.E. They include A as an independent variable, and closely resemble the heat balance equation. Therefore, although it is recognised that there is collinearity due to derivation, the correlation between the values of the energy fluxes appears to be close during this season.

The situation for the autumn months of March, April and May, 1970 is quite different from that of the summer, as can be seen from Table 7.4. Here, although there is still a good correlation between Rn and LE, and Rn and P, all the values of C.C. are lower than their counterparts in the analysis of the annual data. Values of C.C. involving A are particularly low, and A is excluded as an independent variable in the multiple regression equations. When A is the dependent variable, other variables are not accepted into the regression at the 95% confidence level. The simple correlation between LE and P is not as low as in the spring months, but is lower than that given in the analysis of all data. The multiple regression equations have lower values of M.C.C. than the other seasonal regressions, and have much larger values of S.E.E. A

TABLE 7.3

RESULTS OF MULTIPLE REGRESSION ANALYSIS OF DAILY VALUES OF THE HEAT BALANCE COMPONENTS FOR DECEMBER 1969, JANUARY AND FEBRUARY 1970

Correlation Matrix

	Rn	A	LE	P
Rn	1.000	0.624	0.950	0.906
A		1.000	0.549	0.463
LE			1.000	0.744
P				1.000

Multiple Regression Equations

Equation	M.C.C.	S.E.E.
Rn = -0.241 + 0.9999 LE + 0.9998 P + 1.0036 A	1.000	0.630
A = -0.0258 + 0.9911 Rn - 0.9907 P - 0.9905 LE	0.998	0.626
LE = 0.0441 + 0.9997 Rn - 0.9992 P - 1.0028 A	1.000	0.630
P = 0.0356 + 0.9996 Rn - 0.9992 LE - 1.0031 A	1.000	0.630

TABLE 7.4

RESULTS OF MULTIPLE REGRESSION ANALYSIS OF DAILY VALUES OF HEAT BALANCE COMPONENTS FOR MARCH, APRIL AND MAY, 1970

Correlation Matrix

	Rn	A	LE	P
Rn	1.000	0.152	0.887	0.779
A		1.000	0.003	0.159
LE			1.000	0.409
P				1.000

Multiple Regression Equations

		Equation	C.C.	S.E.E.
Rn	=	$-4.8329 + 0.9918\ LE + 1.0291\ P$	0.997	4.603
A		No independent variables accepted at 95% confidence level		
LE	=	$5.3790 + 0.9950\ Rn - 1.0164\ P$	0.995	4.610
P	=	$4.8872 + 0.9335\ Rn - 0.9335\ LE$	0.990	4.419

TABLE 7.5

RESULTS OF MULTIPLE REGRESSION ANALYSIS OF DAILY VALUES OF THE HEAT BALANCE COMPONENTS FOR DATA OF JUNE, JULY AND PART OF AUGUST 1970

Correlation Matrix

	Rn	A	LE	P
Rn	1.000	-0.504	0.913	0.480
A		1.000	-0.541	-0.329
LE			1.000	0.101
P				1.000

Multiple Regression Equations

Equation	M.C.C.	S.E.E.
$Rn = -0.1010 + 0.9994\ LE + 1.0028\ P + 1.0055\ A$	1.000	0.074
$A = 0.0998 - 0.9935\ LE - 0.9969\ P + 0.9941\ Rn$	1.000	0.074
$LE = 0.1012 + 1.0006\ Rn - 1.0034\ P - 1.0061\ A$	1.000	0.075
$P = 0.1007 + 0.9971\ Rn - 0.9966\ LE - 1.0027\ A$	1.000	0.074

relatively large constant term, as well as the absence of A, make the regression equations dissimilar to the heat balance equation.

The reasons for these results may be related to the following factors that apply during this period. Firstly, there are days in April and May when P is directed toward the surface (Appendix F). Secondly, there is a marked difference in the relative magnitudes of daily values of LE and P during the months of April, on the one hand, and March and May, on the other (Fig 7.6). Thirdly, there is a long transition period during which A is not directed consistently either towards, or from, the surface. Also, absolute values of A are relatively low. The lower degree of correlation between the values of the heat flows for this period is interpreted as indicating that the interaction between the flows often occurs in an opposite numerical, and thus physical, direction. This may be regarded as a characteristic of the energy flows for this period.

The direction of energy flows is also important in explaining the results of the multiple regression analysis for the winter months of June, July, and part of August 1970 (Table 7.5). In the simple correlation coefficients, A is inversely correlated with Rn and LE. This is because, on many days during the period, relatively large flows of A towards the surface are associated with small (or small negative) values of Rn, and small values of LE. There is also an inverse correlation between A and P for the same reason, but this is not so strong since P itself is directed towards the surface for over half of the days in the period. The frequent reversal of direction of P also gives rise to its low correlations with Rn and LE. The only really strong correlation in this period remains that between Rn and LE.

Although three of the energy flows often change direction during these winter months, the similarity between the multiple regression equations and the heat balance equation is well marked. The regression equations have high values of M.C.C. and low values of S.E.E. An interesting feature of the equation with A as the dependent variable is that LE and P are selected ahead of Rn as being influential factors. This points to a decrease in the importance of Rn as a primary control during this season, owing to its low absolute values. However, it is still selected first in determining values of LE and P.

Several points arise from the above analyses;

1. Correlation between the daily values appears to exist over and above that due to collinearity owing to the methods of derivation.

2. Some relationships are stronger than others. The simple correlation coefficients for all the data (Fig. 7.2) may be used as a guide to the relative strengths of the relations between the flows.

3. The relative strengths of the relations, as judged by the values of the C.C., vary on a seasonal basis. It is possible that part of the low statistical simple correlation, in the autumn and winter cases, is due to the change of direction of Rn, P, and A as the energy exchange equation (1.1.1.) becomes balanced.

4. Rn is usually most closely correlated to other flows, but

the relationships are not always accompanied by perfect values of C.C. This indicates that factors other than Rn affect the values of LE, P and A. For example, such factors as soil moisture, cloud cover, and humidity, are related to the values of the non-radiative fluxes in different ways. These fluxes may also influence the value of Rn. For instance, flows of P and A towards the surface in winter can markedly affect the value of Rn by allowing a higher value of LW↑ than would otherwise have occurred. It is of interest that such feedback mechanisms, which are not only confined to processes involving Rn, are not encompassed by climatonomic theory in the form used earlier.

5. Generally low correlations exist between A and the other flows. Although possibly due to instrumental and sampling error, another factor may be the relatively slow passage of heat through the soil compared with that through air. A relation between A and the other fluxes most probably exists (see section 3.4), but it has been found frequently in analyses (e.g. section 6.2) that A acts in a way that is, in some respects, different to that of the other main flows. In other locations, results supporting this conclusion have been reported. Vowinkel (1966), for example, found that for Ottawa and Resolute, soil heat storage changes were affected predominantly by seasonal variation rather than by the synoptic scale weather.

7.3 The Surface Heat Balance and Synoptic Weather

7.3.a. Rationale

This section is concerned with the effect of synoptic weather conditions on the values of the surface heat balance. The direction of cause and effect implied here is governed by the restriction of the study to the heat balance at one location. It should be mentioned that the surface heat balance, when operating over a large area, can affect the synoptic weather (Daigot and Sakali, 1969), although the details of the interaction are by no means clear (Sheppard, 1969). The purpose of the present examination is to show that the addition of information on the energy fluxes helps to give a more complete description of the climate, than would be achieved by an exclusive discussion of synoptic variations, and their effect on standard climatic parameters. An examination of the relation between synoptic weather and the heat balance, with emphasis on values of A and the effect of rain, was given in section 3.5. In the present section a more systematic examination of the effects of the most frequent kinds of synoptic situations is attempted.

Watts (1947), using a sample of 1096 days, has shown that for most of the South Island, New Zealand, four main types of synoptic situation prevail for a total of 76% of the time. These are:- (1) anticyclones (20%), (2) moderate or strong northwesterly or northerly flow (15%), (3) cold fronts (10%), and (4) weather at the rear of cold fronts (31%), for which a southwesterly air flow is the most common. The effects of each of these situations will be examined with respect to daily values of the surface heat balance (Appendix F, and Figs. 7.3.a. and 7.3.b.).

7.3.b. Anticyclonic Situations

The best example of anticyclonic weather during the study period was between 30 October and 20 November 1969, when a blocking high pressure zone, described by Hill (1971), was present over almost all of New Zealand. The only interruptions occurred on 10 and 12 November when weak frontal systems affected the study area, but these gave no precipitation. The effect of the anticyclone on the surface energy flows is seen in Fig. 7.3.a. At the beginning of the period Rn was consistently high. Values of LE and P were also high, but, on a day to day basis, fluctuated more than the values of Rn. The Bowen ratio was approximately 0.50. There was a relatively high soil heat flow away from the surface. The passage of the weak frontal systems divided the anticyclonic period into two parts. In the second part, values of LE rapidly decreased and then remained relatively low, owing to a lack of soil moisture. Values of P became high and Rn was also large. This latter part of the period was unique in the study year, because for ten days turbulent flow from the surface was predominantly of sensible heat. Bowen ratio values were well above unity. The highest recorded was 6.1 which is comparable to those found in dry seasons in tropical climates (Polavarapu, 1968). Throughout the whole period of 30 October to 20 November, daily soil heat flow values showed flows into the soil ranging between -4 and -27 ly day^{-1}.

An analysis of surface weather charts for 0000 hr (N.Z.M.S., 1969-70) showed four other clearly marked anticyclonic periods in the study year. In this case, an anticyclonic period is defined as a period when the South Island is in, or near, the centre of closed, high pressure isobars; no front reaches

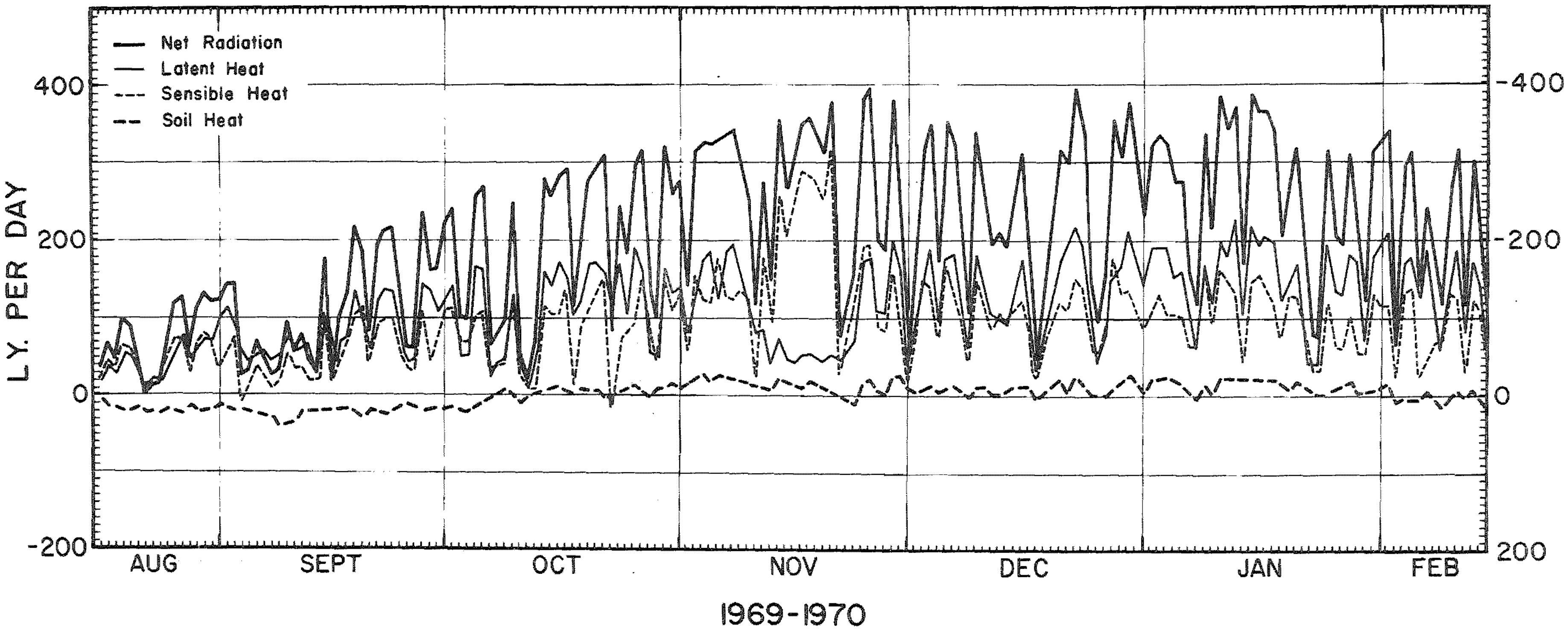
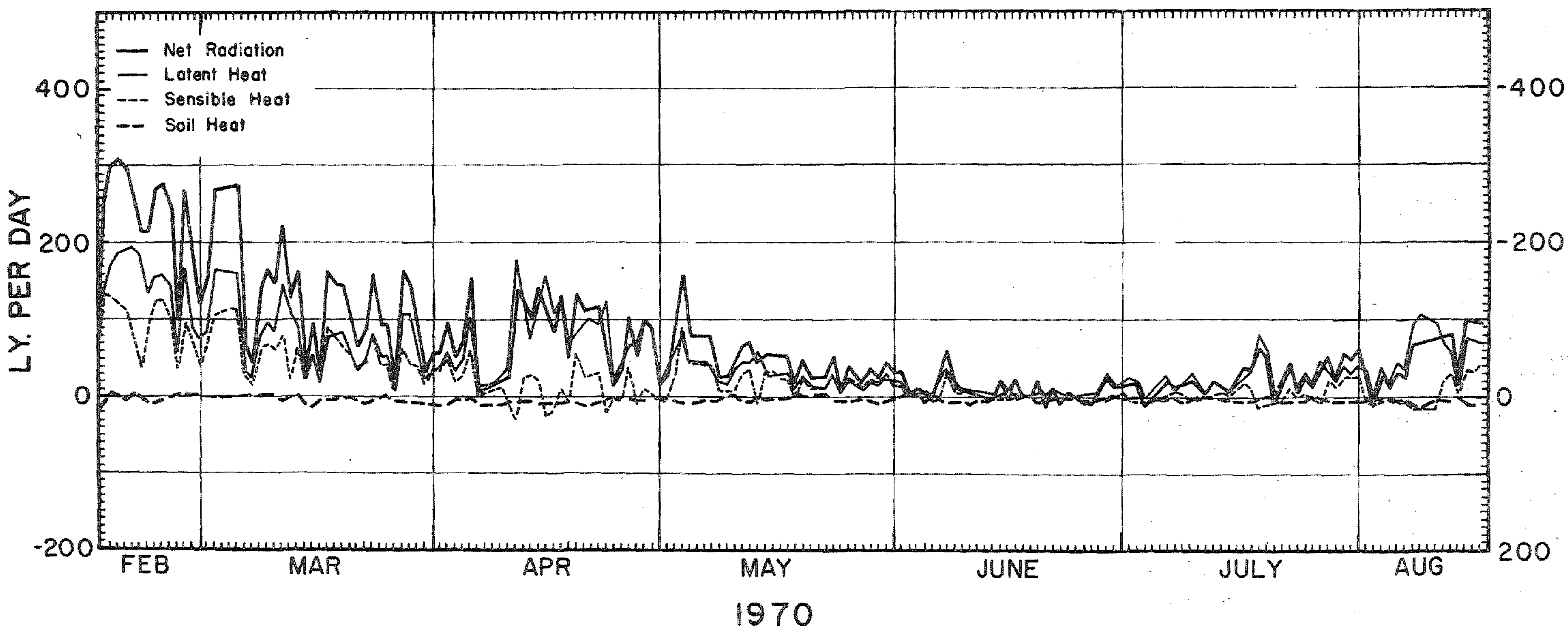


FIGURE 7.3.a: Daily values of the heat balance components at the Chilton Valley during the first part of the study year. (Left hand scale relates to values of Rn; right hand scale relates to values of LE, P and A)

FIGURE 7.3.b: Daily values of the heat balance components during the second part of the study year. (Left hand scale relates to values of Rn; right hand scale relates to values of LE, P and A)



any part of the land surface, and the situation persists for four or more days. The periods were 23-26 August 1969, 12-15 October 1969, 14-18 February 1970, and 28 May to 2 June 1970. All of these periods showed similarities, in the relative values of energy flows, to the first part of the period described above. The value of Rn was usually high for the time of year. The October and February cases showed Bowen ratios of approximately 0.6, although those of winter were about 1.0. The direction of soil heat flow depended on the season, only the February case showing no definite pattern.

On the basis of this evidence, it appears that persistent anticyclonic weather may be accompanied, in the Chilton Valley, by high values of Rn. This energy is dissipated by LE and P in about equal proportions in winter, with loss by LE being rather larger, than that by P, in other seasons. The flow of soil heat is dependent mainly on the time of year. The last half of the period 30 October to 20 November showed that if the weather type is very persistent, at least in the warmer months, then P can have very high values with respect to LE.

7.3.c. <u>Moderate to Strong Northwesterly or Northerly Flow</u>

Winds from the northwest or north have been shown (section 5.3) to be the most frequent in the general locality of the Chilton Valley. Moderate to strong winds from these directions, although not exclusively, are usually associated with a synoptic situation of the kind shown in Fig. 7.4, which is for 0000 hr on 24 December 1969. Moderate to strong northwesterly or northerly flow is given by the western side of an anticyclone located to the east of New Zealand. It is also quite common for a cold front, associated with an eastward moving trough

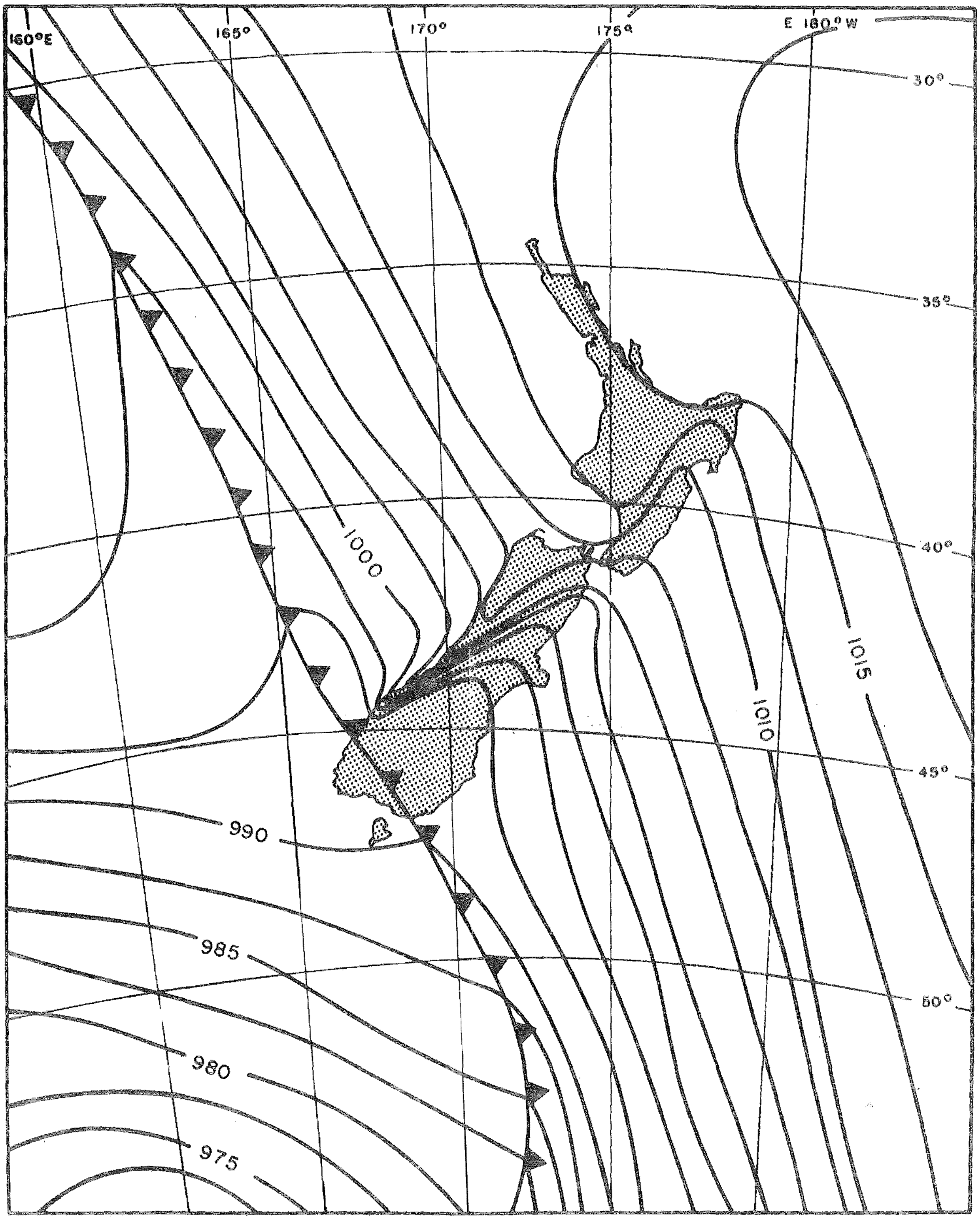


FIGURE 7.4: Synoptic situation for 0000 hr on 24 December 1969 illustrating conditions leading to a moderate to strong northwesterly or northerly flow

of low pressure, to be present to the south west of the country (Lamb, 1970). Days during the study period, when the synoptic situation was similar to that shown in Fig. 7.4, with reference especially to the isobaric pattern over the South Island, are listed in Table 7.6. The heat balance for the appropriate days is also shown.

Three interesting points arise from the values in Table 7.6. Firstly, the value of Rn during this weather type was not always high since SW↓ was frequently decreased owing to cloud cover. The cloud cover might have been due to orographic effects (sometimes leading to the formation of a föhn cloud), or to cloud associated with the approaching cold front. Occasions when north or northwesterly air flow persisted for a second successive day, were, more often that not, associated with lower Rn values on the second day. A second point is that the flows of LE away from the surface were always greater or equal to those of P. In summer, the value of LE could be very much higher, as for example, in the case of 6 February 1970. This was due to the föhn conditions that developed with their associated high wind speeds and low humidities. These conditions have been discussed in sections 4.4 and 5.5. High values of LE have also been found on the Canterbury Plains (Lamb, 1970), and in other locations where föhn winds are common, such as Alberta (Ashwell, 1967; Marsh, 1965). Table 7.6 shows that in September 1969, and April to August 1970, under a north or northwesterly air flow, P was directed towards the surface. This flux supported values of LE that sometimes exceeded those of Rn. Finally, when these synoptic situations occurred, the values of A, and especially the direction of the flow, were

TABLE 7.6

HEAT BALANCE VALUES IN LY DAY^{-1} FOR DAYS WITH MODERATE OR STRONG NORTHWESTERLY OR NORTHERLY AIR FLOW

Date	Rn	A	LE	P
3.9.69	25	18	-52	9
7.9.69	25	28	-45	-7
15.9.69	22	21	-25	-18
24.9.69	113	15	-71	-57
5.10.69	269	3	-164	-108
7.10.69	81	-2	-43	-36
30.11.69	50	-8	-25	-17
7.12.69	231	-5	-128	-99
8.12.69	101	2	-60	-43
15.12.69	310	-11	-177	-122
21.12.69	298	-6	-186	-107
23.12.69	350	-18	-195	-138
24.12.69	147	-3	-84	-59
13.1.70	163	-23	-99	-40
20.1.70	316	-19	-172	-126
25.1.70	206	-11	-134	-62
26.1.70	194	-9	-128	-57
6.2.70	241	-5	-190	-46
10.2.70	316	-6	-188	-122
11.2.70	102	6	-79	-29
6.3.70	66	-4	-35	-27
9.3.70	166	-3	-97	-67
20.3.70	109	1	-55	-55
24.3.70	92	1	-52	-42
2.4.70	97	11	-58	-51
24.4.70	15	0	-17	3
25.4.70	34	-1	-40	7
17.5.70	53	2	-32	-22
4.6.70	6	6	-8	-5
5.6.70	-4	-1	0	5
14.6.70	20	-4	-19	3
23.6.70	4	0	-4	1
28.6.70	26	-1	-30	5
10.7.70	15	-1	-17	3
18.7.70	64	3	-80	13
19.7.70	50	0	-61	10
25.7.70	14	-1	-16	3
26.7.70	38	2	-47	8
1.8.70	23	6	-34	6
2.8.70	-11	1	0	10
3.8.70	28	4	-39	6
13.8.70	25	1	-18	-7

usually influenced by the season of the year. In the colder months, soil heat flow towards the surface helped to support a relatively high value of LE, as in the cases of September 1969.

With regard to climatic explanation, the attention drawn by the heat balance approach to the relatively low values of Rn, high values of LE, and the downward flow of P in winter, is vital to an understanding of the effect, on the climate, of north or northwesterly flow, itself so common in the location being studied.

7.3.d. Cold Fronts and Successive Southwesterly Flow

The weather accompanying a cold frontal passage is generally fresh to strong northwesterly winds and clouds, followed by fresh and occasionally strong winds from a southerly quarter. Usually the actual passage is marked by a short period of moderate or occasionally heavy rain, followed by colder, showery conditions (Maunder, 1970). Within this framework, the actual weather associated with any one cold frontal passage is variable. It depends, in any one area, mainly on the direction of airflow that follows the front. Only selected cases, where principally southwesterly flow follows the front or fronts, are examined below. Even in these cases there is far more variability of heat balance response than was found in the synoptic situations investigated above. When it is realised that a cold front or fronts were located over New Zealand on more than 45 days during the study period, and that variations of any one of five types of following air flow (Watts, 1947) could occur, then it becomes apparent that a high degree of variability of cold frontal weather and response might be expected to have occurred.

The limited number of cases studied (Table 7.7) showed that southwesterly weather following a cold front, could itself be caused by two different synoptic situations. The first, designated type I, occurred in southwesterly flow on the eastern side of an anticyclone (Fig. 7.5). This type was found also on the south side of the high pressure area, in which case, the flow had a greater westerly component. The second situation (Type II) was that where the front was in a trough of low pressure approaching the southwestern side of an anticyclone (Fig. 7.4). The cases listed in Table 7.7 include both types of situation, and show values of the heat balance, air temperature, and precipitation for the days before the front, the day or days of the front, and the days when south-westerly air flow followed.

The values in Table 7.7 show that for both types I and II, there was no set pattern followed by the values of the energy fluxes on days characterised by this weather. The main reason for this may be the variety of factors that affect the value of Rn. Although in many cases the value of Rn was relatively low on the day of the front, as might have been expected due to the effect of cloud cover on SW↓, cases 4, 6, 8 and 10 were exceptions to this. Similarly, the values of Rn on the days following the frontal passage could be either high (case 2) or low (case 11). The values of LE and P were also variable, but tended to follow the trend set by those of Rn in any particular case. There was a tendency for Bowen ratios to have been below unity on the day preceding the front, and to

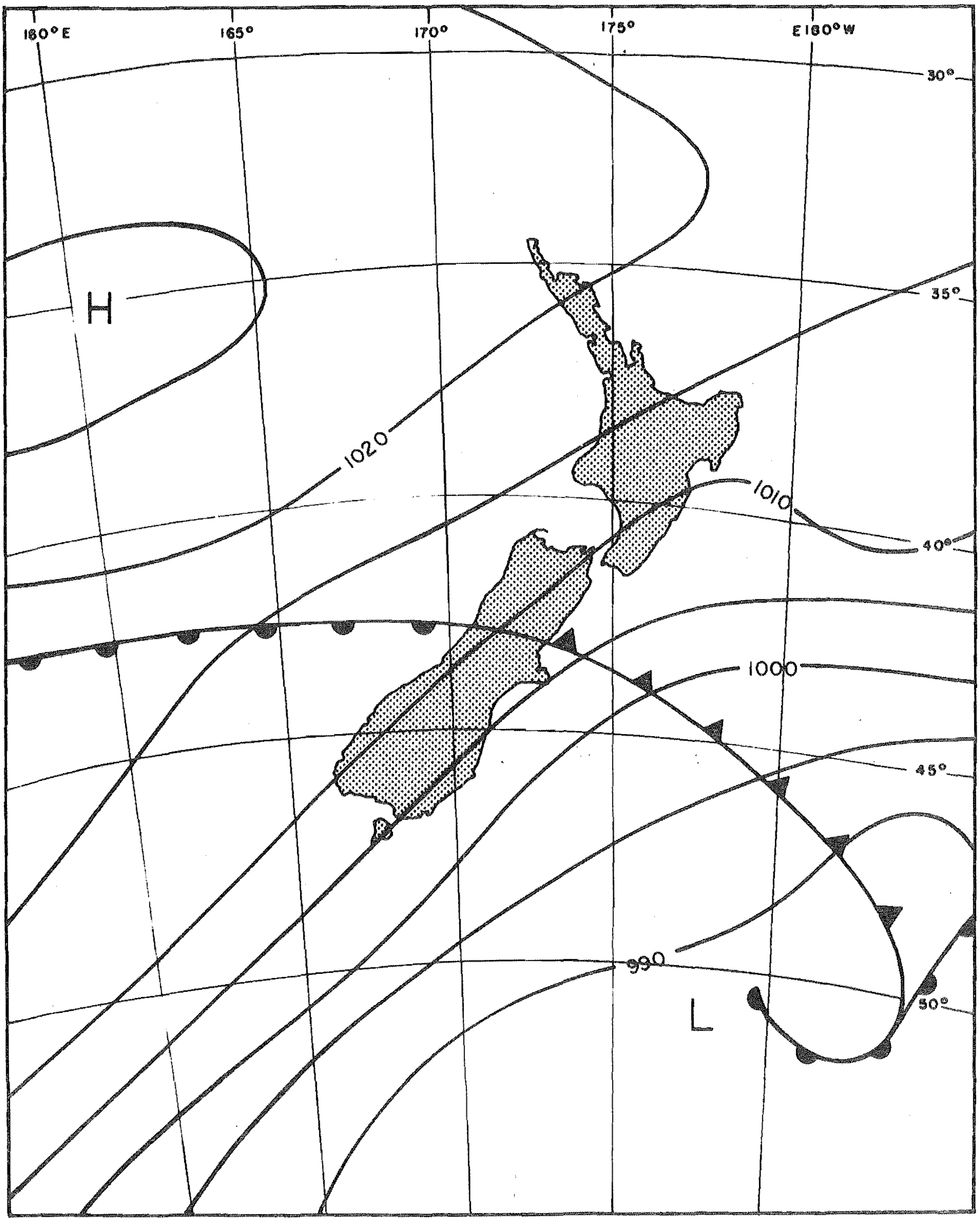


FIGURE 7.5: Synoptic situation for 0000 hr on 19 April 1970 illustrating a southerly front and successive southwesterly flow

TABLE 7.7

HEAT BALANCE VALUES IN LY DAY^{-1}, MEAN AIR TEMPERATURE AND PRECIPITATION FOR DAYS DURING PASSAGE OF COLD FRONT OR FRONTS. DAY OF PASSAGE OF FRONT MARKED BY ASTERISK

Case No.	Date	Type (see Text)	Rn	A	LE	P	Temp. °C	Precip. mm
1	19.8.69	I	92	+21	-52	-61	2.7	0
	20.8.69*		58	14	-38	-34	0.7	1.9
	21.8.69*		1	22	-11	-13	-1.3	3.8
	22.8.69		13	20	-22	-11	1.3	0
2	26.8.69	I	127	22	-76	-73	4.0	0
	27.8.69*		61	13	-47	-27	3.0	3.9
	28.8.69		112	22	-65	-68	3.6	0
3	1.10.69	I	242	14	-142	-114	8.7	0
	2.10.69*		101	20	-52	-69	3.5	0.5
	3.10.69		95	22	-50	-67	0.7	0
4	7.10.69	II	81	-2	-43	-36	6.1	2.5
	8.10.69*		98	-8	-47	-42	7.0	0.1
	9.10.69*		248	-1	-129	-118	5.2	2.2
	10.10.69		51	11	-37	-26	-2.0	5.0
	11.10.69		17	0	-9	-8	1.8	0
5	23.12.70	II	350	-18	-195	-138	15.7	1.4
	24.12.70*		147	-3	-84	-59	15.9	11.8
	25.12.70*		93	0	-45	-48	10.4	72.2
	26.12.70		161	0	-77	-84	9.2	11.4
6	1.2.70	II	340	-14	-210	-116	15.5	0
	2.2.70*		72	10	-64	-16	9.1	0
	3.2.70*		295	4	-168	-131	11.7	3.1
7	6.2.70	II	241	-5	-190	-46	19.1	0
	7.2.70*		177	0	-110	-67	16.2	3.6
	8.2.70		112	16	-61	-68	4.3	12.6
8	18.4.70	I	49	6	-66	+11	14.0	0
	19.4.70*		132	8	-85	-55	8.6	0.4
	20.4.70		110	12	-96	-26	5.1	0
	21.4.70		117	11	-101	-27	7.1	0
9	12.5.70	I	71	8	-44	-35	2.0	0
	13.5.70*		47	0	-57	+9	7.0	0
	14.5.70		56	5	-32	-29	3.5	0
10	22.6.70	I	-8	5	0	3	1.5	0
	23.6.70*		4	0	-4	1	6.5	1.9
	24.6.70		-1	2	-1	0	5.8	0
	25.6.70		-9	4	0	5	4.4	0
11	22.7.70	I	45	6	-42	-9	3.0	0
	23.7.70*		6	4	-12	+2	3.4	0
	24.7.70		26	5	-28	-3	3.0	0

have risen to above unity in the later southwesterly weather (cases 2, 3, 5 and 7), but this was not seen in the values of the other examples. There was a different reaction of soil heat flow to weather types I and II. In the latter, flow into the soil showed a tendency to be reversed during the three to five days of the synoptic weather situation. The magnitude of the flow was not closely related to the value of Rn. In cases of flow type I, flow to the surface usually was decreased in size on the day of the front (cases 1, 2, 9, 10 and 11). The reason for this is not clear. It can sometimes be explained in terms of increased mean air temperatures (cases 9, 10 and 11). It also might have been due to precipitation entering the soil (cases 1 and 2), although the latter possibility is not supported by the evidence in section 3.4.

Therefore, it can be seen that unlike the cases of anti-cyclonic and northwest to northerly air flow situations, it is difficult to generalise on the heat balance response to the atmopsheric conditions accompanying cold frontal passages in this location. This is true, even when the latter are limited to types with a following southwesterly air flow. Often, variability of cold frontal weather types is emphasised in descriptions of the climate of New Zealand (e.g. Watts, 1947; Maunder, 1970). The variation of the heat balance responses in the Chilton Valley justifies this emphasis, and adds depth to these descriptions.

7.4 The Annual Heat Balance

The mean monthly values of the heat balance for the

Chilton Valley during the study year are shown in Fig. 7.6. It is seen that Rn had the highest values at all times, except for a short period at the end of the study year. In the monthly mean values Rn did not become negative in the winter. The input of heat energy by Rn was dissipated mainly by the evaporative and sensible heat flows. The former was usually the larger heat loss, except for the notable period in November, when there was a lack of soil moisture. Monthly mean values of P did not show a net flow to the surface in winter. Soil heat flow was small in comparison with other flows. However, a small flow of soil heat to the surface at the end of the study period was partly responsible for values of LE exceeding those of Rn.

A comparison with the heat balances in other locations outside New Zealand is made in Table 7.8 and Fig. 7.7. Three difficulties arise in making a direct comparison. Firstly, the other heat balances are for different years. Secondly, the likely magnitude of error is not always quoted in the works from which the other heat balances are drawn. Where errors are quoted (Budyko, 1958), except for those in P, they are of the same order as those believed to exist in the present study (section 1.5). Thirdly, the microclimatic conditions existing at the other stations vary at each station, and are different from those at the Chilton Valley. Despite these difficulties, the broader scale aspects of the heat balances at the different stations may be compared with advantage. All of the stations are selected because they may be used to illustrate certain noteworthy features of the heat balance at the Chilton Valley. The stations, with the exception of

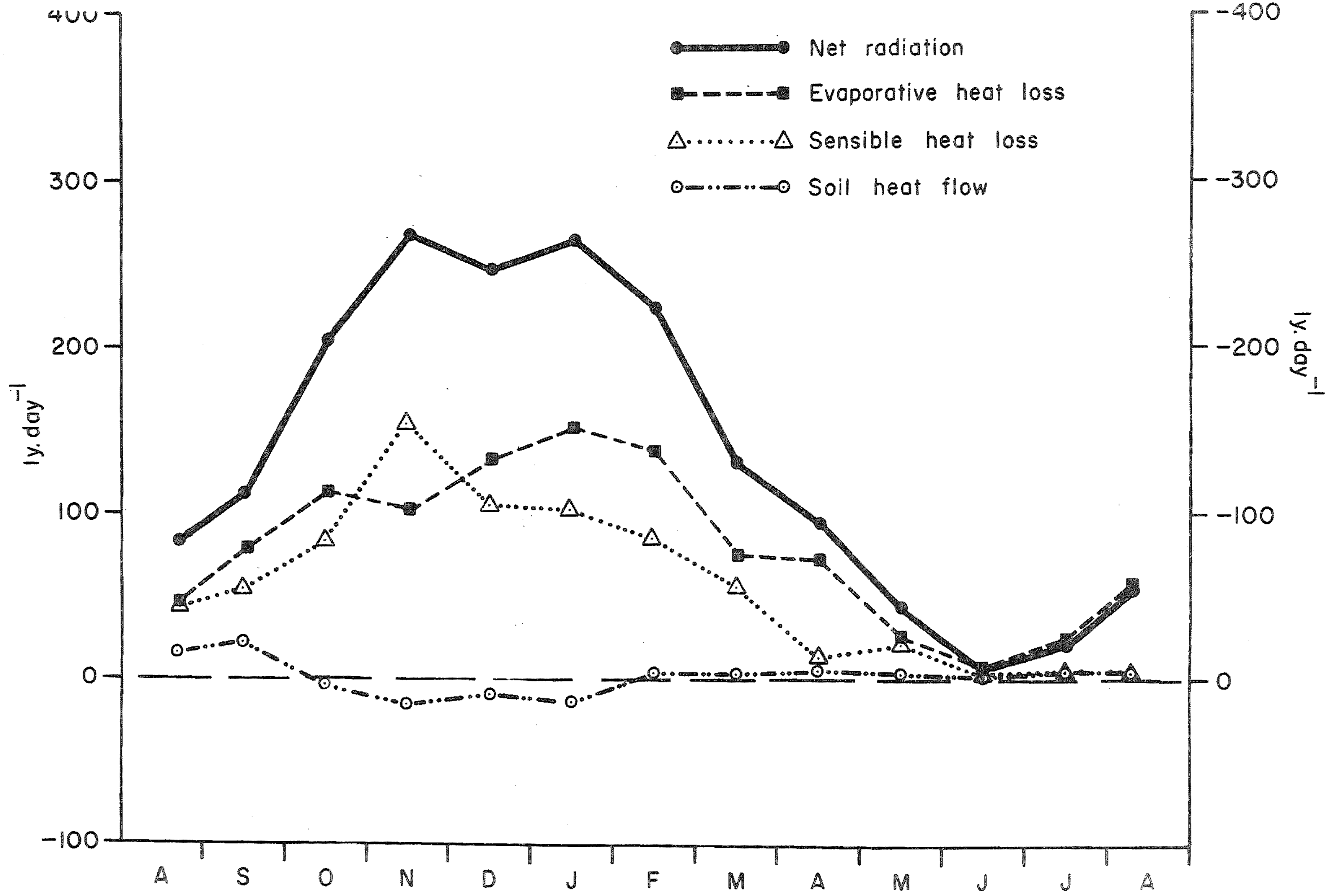


FIGURE 7.6: Monthly mean values of heat balance components at the Chilton Valley during the study year. (Left hand scale relates to values of Rn and A; right hand scale relates to values of LE and P)

TABLE 7.8

COMPARISON OF ANNUAL HEAT BALANCES. HEAT BALANCE VALUES IN LY DAY^{-1}

Location	Latitude	Date	Source	Rn	A	LE	P
Chilton Valley	43 °S	Aug 1969 Aug 1970	-	136	2	-82	-58
Quickborn W.Germany	54 °N	July 1957 June 1958	Frankenberger (1962)	105	8	-86	-20
Paris France	49 °N	Climate Data	Budyko (1958)	95	-	-72	-24
Lisbon Portugal	39°N	Climate Data	Budyko (1958)	151	-	-86	-67
Astoria Oregon	46 °N	Climate Data	Sellers (1965)	156	0	-99	-58
Saigon Vietnam	11 °N	Climate Data	Budyko (1958)	213	-	-156	-59
Fairbanks Valley Alaska	65°N	Aug 1966 July 1967	Wendler (1970)	61	-4	-74	17
Fairbanks Hill Alaska	65°N	Aug 1966 July 1967	Wendler (1970)	24	-3	-19	-2

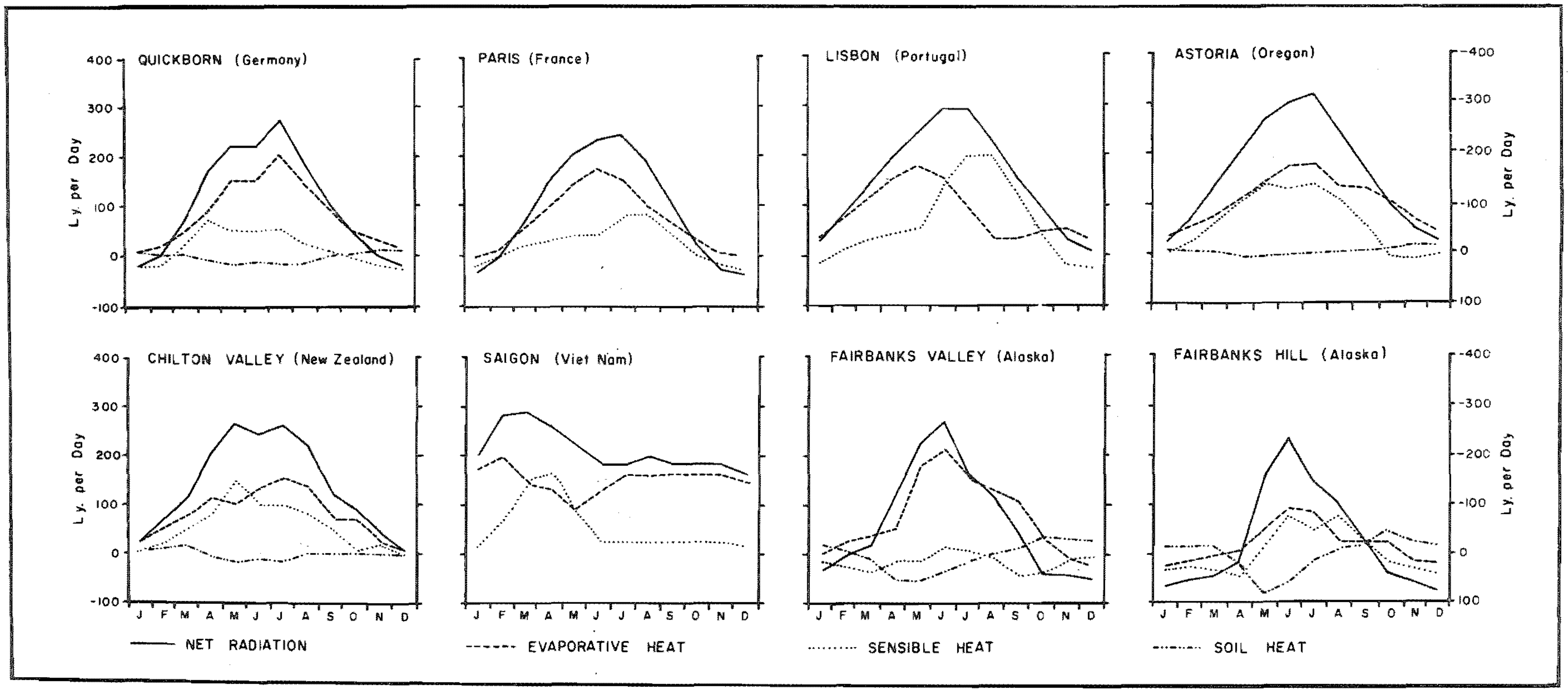


FIGURE 7.7: A comparison between the heat balance estimates at the Chilton Valley during the study year, and estimates from other world stations. Data sources are Budyko (1958), Sellers (1965), Frankenberger (1962) and Wendler (1970). (Left hand scale relates to values of Rn and A; right hand scale relates to values of LE and P)

Saigon, and possibly Lisbon, are near the west coasts of land masses, and are affected by a maritime influence which, in general, is associated with westerly winds.

In the annual data (Table 7.8) it is seen that, except for Fairbanks (at higher latitude) and Saigon (at lower latitude), the value of Rn is comparable to that found at the other stations. The soil heat flow values, although assumed to be zero by Budyko (1958), are seen to have small finite values at other stations (Fairbanks and Quickborn) where estimates are based on the data of only one year. The Chilton Valley LE value is comparable to those of the western European sites. It might well have been as high as that of Astoria if adequate soil moisture had been available during all of the year. The value of P found for the Chilton Valley is more comparable to those of Astoria and Lisbon and even Saigon (where most of the sensible heat loss occurs during a short, dry period), than to those of Quickborn and Paris, where a much higher proportion of heat is lost by evaporation. Overall, the annual heat balance for the Chilton Valley is similar to that of other mid-latitude sites near the west coast of land masses. The altitude of the valley does not make the heat balance at all similar to that of either the hill or valley stations of Fairbanks at higher latitude. However, the amount of sensible heat transfer is similar to that found at a tropical station which experiences a dry season.

In the variation of the heat balances through the year (Fig. 7.7) it is interesting to note that Rn, at the Chilton Valley, did not become negative in the winter, whereas net radiant flow away from the surface occurs at some of the other stations (Quickborn, Paris, and Fairbanks). The reasons for the absence of a negative net radiation balance in the Chilton Valley may include the following. Firstly, there may be a relatively increased amount of LW↓ in the valley due to the presence of the surrounding hills. Secondly, there may be also a relatively increased value of SW↓, compared with the other stations, owing to less cloud cover and/or higher atmospheric transmissivities due to the effect of altitude. As mentioned in section 2.6, Samukashvili (1969a) working at an altitude of 550 m and at latitude 43^{o}N, also found monthly mean values of Rn to be generally positive throughout the year. Overall, the absolute magnitude of the annual variation of monthly mean values of Rn at the Chilton Valley is quite comparable to that at most of the other stations. Small differences in the slope and orientation of the surface, (section 2.9) and/or errors in estimation (section 1.5), could account for all of the variations of Rn between the Chilton Valley and the other mid-latitude stations.

A discussion of the direction of flow is also relevant for the sensible heat flux. In this case, Fig. 7.7 shows that the Chilton Valley is the only station, with the exception of Saigon, not to exhibit a net flow of sensible heat towards the surface in winter. This was partly due to the relatively high values of Rn in this season, and possibly, partly due to (1) the method of computing LE which used the value of Rn, and (2) then computing P as the residual of the heat balance. However, the many individual days (see Fig. 7.3.b., and Appendix F) when both Rn and P were directed towards the surface, to support a value of LE exceeding that of Rn, suggest that the value of Rn rather than the method of computing P, may have been the most important cause. It should also be mentioned that the possibility of net downward flow of P in the winter exists within the range of error quoted for the monthly mean values (section 1.5).

Apart from the lack of positive values of P in winter, the P values at the Chilton Valley resembled those of Astoria, more than those of Paris or Quickborn. The latter two have less negative values in the warmer months than did the Chilton Valley. It is of interest to note that the relatively high negative value of P, at the Chilton Valley in November, was closer to that of the dry season for Saigon, in March and April, rather than that of the dry season of Lisbon, in June and August. The inference to be drawn from this observation is that the climate of the Chilton Valley is better described as a humid climate, that can have a short dry season, rather than a dry climate that has a long humid period.

The values of LE for the Chilton Valley are more in agreement with those of Paris, Quickborn and Astoria, which have generally humid climates, and where LE appears to be determined mainly by the value of Rn. This again supports the above inference. The relatively low LE value for the Chilton Valley however, does represent a marked difference from the other mid-latitude examples. In these, with the exception of Lisbon, loss of latent heat does not appear to be influenced, on a monthly basis, by a lack of soil moisture. A further difference is that the evaporative heat loss in the winter in the Chilton Valley is less than that in the other mid-latitude stations (except for Paris). This may be due to the absence of net downward sensible heat flow and may be associated with the altitude of the station. Also, the long period of shading by sky line obstructions will lower the value of LE in winter. It must be remembered that the Chilton Valley estimates are of lowest accuracy in this season, a fact that applies to all of the components discussed so far in this section. Although the latent heat loss was relatively low in this season, the model used does not allow the flow to be directed towards the surface, as can happen (e.g. Fairbanks Hill). This is not believed to introduce a large error in the present case.

The soil heat flow values for the Chilton Valley are in the same order as those published for Quickborn and Astoria, and are typically very much less, in absolute magnitude, than those of the other components. In the Fairbanks examples, the magnitudes of A are high mainly because heat involved in melting and freezing surface water is incorporated into values of A.

The most important feature that emerges from these comparisons is that the heat balance of the Chilton Valley during the study year was broadly the same as that of sites located at similar latitudes, and in similar positions with regard to maritime influence. Apart from the general similarity, the following differences for the study year were noteworthy. Firstly, Rn did not become negative in winter. Secondly, P did not show a net downward flow in winter. Finally, although LE was approximately the same as that at other mid-latitude stations, for most of the year, soil moisture deficit in some parts of the year markedly affected its value, and consequently the values of P.

7.5 Summary

The main points arising from these studies of the values of the heat balance at the Chilton Valley are as follows:-

1. Relationships appeared to exist between the daily values of the energy flows.
2. The relationships between some flows, especially those involving Rn, were stronger than others, such as those concerned with A.
3. The relative strengths of the relationships varied on a seasonal basis.
4. Similarities in the relative values of heat balance response were seen in anticyclonic and northwesterly or northerly air flow situations. However, the heat balance response to synoptic situations involving the passage of a cold front, and the successive southwesterly air flow, was variable.
5. The heat balance of the Chilton Valley was broadly the same as that of sites located at similar latitudes, and in similar positions, with regard to maritime influences and the effect of the westerlies.
6. Apart from the general similarity, three notable differences were; the positive and negative mean monthly values respectively, of Rn and P, in winter, and the high Bowen ratio values of November.

CHAPTER EIGHT

CONCLUSION

8.1 The Character of the Climate of the Chilton Valley

8.1.a. Introduction

The principal aim of this study has been to demonstrate that a knowledge of the climatic energy exchanges gives an insight into the character of the climate of the Chilton Valley. In section 1.4, the climate of the location was described, for the most part, in terms of standard climatic parameters. Some aspects of the character of the climate were implicit in this description. It is the purpose of this section to show explicitly the advantages gained, and the consequent insight given, as a result of the heat balance approach. A better understanding of, and more information on, the parameters in section 1.4 has been given. Furthermore, the benefit of the knowledge of the energy flows themselves, is available. These two aspects will be examined. The insight gained into the character of the climate of the Chilton Valley will then be clear.

8.1.b. <u>The Climate in Terms of the Standard Climatic Parameters</u>

The climatic parameters used in section 1.4 are temperature, precipitation, wind, humidity and shortwave radiation. Factors arising from the heat balance study, that are relevant to each of these parameters, are discussed below.

The air temperature measured by a standard Stevenson screen integrates the temperatures of the multitude of parcels of air near the earth's surface and the air aloft (Miller, 1966). Viewed in this way, screen temperature is only an indirect index of part of the environment. It is dependent upon all the flows of energy in the heat balance equation (equation 1.1.1.), and, in particular, on Rn and P. It can also be affected by advection. Two examples from the present study illustrate this dependence.

Firstly, the high mean and maximum temperatures of the summer months (Tables 1.1, 1.2, 6.1) are principally a result of the following energy exchange factors. Owing to the comparatively low albedo, a large proportion of the relatively high SW$\downarrow$ is absorbed at the surface. Although some of the radiant energy is lost as LW$\uparrow$, the resulting Rn is higher than it would have been if α were larger. More heat is therefore available to warm the air and the submedium. This is especially so when evaporative heat loss is lowered as a result of the possible summer soil moisture deficit conditions (Chapter 4). Advection (section 5.5) has also been shown to be a possible cause of high summer air temperatures at times when north-westerly airflow prevails.

A second example concerns winter temperatures. The mean and minimum, June, July and August air temperatures (Tables 1.1, 1.2, 6.1) are not as low as might be expected at this altitude in a mountain environment. Evidence of this is given by Morris (1965) who, with reference to the nearby Craigieburn site, shows that winter temperatures are 'markedly dissimilar' from those of otherwise comparable mountain climates in Western Europe and North America. He attributes this to the oceanic influence. De Lisle (1970) has indeed shown that large amounts of sensible heat are withdrawn from the Tasman Sea. The values of P (Chapter 5, Appendix F) demonstrate that in the winter of the study year, presumably as a result of advection, it was quite common for P to be directed towards the surface. However, in addition, the present study has shown that during the winter of the year examined, the mean monthly net radiation balance remained positive. This also would contribute to the resulting, relatively mild, temperatures. Although not directly related to the heat balance approach, evidence has also been given suggesting the existence of a thermal belt (section 5.5) at the study site. If this is confirmed, it may also help to explain the winter temperature characteristics.

Another temperature related feature of the climate of the Chilton Valley is the formation of needle ice. Although not complete, the investigation of this phenomenon (section 3.6) at least indicates a range of values of Rn and A under which needle ice formation is possible in the field, given sufficient moisture. Laboratory experiments demonstrate that the actual flow by which energy is lost from the surface is not necessarily important.

Rainfall data on amounts, temporal distribution and intensities (section 1.4) are very important in elucidating the climatic character of the Chilton Valley. However, the heat balance approach has gone further by demanding data on evaporation, which, in turn, has required a study of several aspects of the water balance. As a result, more data are now available on evapotranspiration, percolation, the possibility of soil moisture deficits, the manner of melting and drying of the soil, and the thermal response of the soil to rainfall. In addition, although little more is known about the frequency of snowfall, a possible effect of snow cover on albedo has been seen for this location. Finally, the application of climatonomic theory has drawn attention to the relatively high loss of rainfall by percolation through the soil.

All of this information represents an increase in our comprehension of the character of the climate. Possibly the most illuminating features are firstly, the fact that evaporative water losses are rather lower than those predicted by either the application of standard empirical formulae, or the adjustment of open pan evaporation values; and secondly, despite this, it is quite possible for soil moisture deficits to be a factor of the climate in the summer months.

An investigation of wind in the study year (sections 5.3 and 5.4) shows that velocities were lower than those found at the Biological Station in an earlier period. The importance of the valley topography in determining wind direction has been quantified. Preliminary data on the important aspects of the wind profile, u_* and z_o, have been obtained. Besides this, the necessity for recognising wind as a factor in determining daily evapotranspiration amounts in this location, has been made (sections 4.3 and 4.4). Little further information on humidities has been obtained but the same sections also show the importance of saturation deficit in the estimation of daily evaporative water loss.

SW↓ data at the Chilton Valley (section 1.4) has been supplemented by an empirical knowledge of the way it is related to Rn. Initial information is also available on the amount of SW↓ reflected from different surfaces at different times, the relative size of SW↓ in relation to the other components of the radiation balance at the site, and the relative magnitudes of Q arriving at other slopes in the valley. Here again an understanding of the character of the climate is enhanced as a result of heat balance or related studies.

8.1.c. The Climate in Terms of the Energy Flows

The present study has given a greater understanding of the character of the climate than is possible from a description based on standard climatic parameters. But it has gone further in providing hitherto unknown data on the energy fluxes themselves. The values of these fluxes were fully discussed and analysed in Chapters 6 and 7. The character of the present climate is illustrated by the following conclusions from earlier chapters.

Daily variability, a cyclic nature, and interrelationships are all apparent in the energy flows. A high day to day variability has been noted in many of the parameters associated with the heat balance components and in the values of the components themselves. A quantitative measure of this is given by the relative importance of high frequencies in the time series analyses of the daily heat balance values. The flows of heat energy exhibit important cycles, that are mainly due to astronomical considerations, but which may also have other causes. The fluxes of heat energy are themselves interrelated but the strength of the interrelationships can vary at different times of the year.

The periodic nature of the energy flows at the Chilton Valley is emphasised by an application of climatonomic theory because of its use of Fourier series. The theory, as applied in the present study, not only illuminates the loss of water by percolation but also shows that a variable Bowen ratio is an important feature of the climate during the study year.

Some synoptic weather systems are shown to affect the values of the heat balance at the Chilton Valley. Similarities of heat balance response are seen in anticyclonic, and north-westerly air flow conditions. However, the passage of a southerly front, and the air flow that follows, can give rise to a variety of energy flow situations.

A comparison with the heat balances of other world stations in broadly similar geographic locations, primarily shows that the heat balance of the Chilton Valley bears a general resemblance to these. However, the possibility of positive mean monthly Rn and negative mean monthly P, in winter, and of high negative P values in summer demonstrates noteworthy characteristics of the heat balance of the Chilton Valley.

8.1.d. Conclusion

The heat balance approach has focused attention on climatic factors that may otherwise have been overlooked. It has added more information to a knowledge of the climate of the location. It has also dealt with causal factors of the climate rather than with the results. Therefore, in several ways, a heat balance exposition of the climate of the location has added a further dimension in understanding. The principal aim of the study has thus been achieved and a knowledge of the climatic energy exchanges has given an insight into the character of the climate of the Chilton Valley.

8.2 Comments on the Spatial and Temporal Extension of the Results of the Present Study

Caution is advised in any attempt to extend the results of the present study on any spatial or temporal scale. In order to relate the heat balance of the Chilton Valley to other mountain areas of the South Island, for example, a further comprehensive and systematic study, such as performed by Frankenberger (1962) for Schleswig-Holstein, would have to be made. However, estimates of heat balances of particular locations are often, at least by implication, taken to be representative of larger scale climatic areas (Budyko, 1958; Gates, 1962). The absence of other heat balance estimates for the South Island High Country makes it possible that other workers will wish to refer to the present study. In this event, the comments listed below should be used to help assess the utility of the results that have been presented.

1. Energy exchanges at surfaces such as forest, bare scree

and rock, snow, river bed, and lakes have not been studied.

2. The longer period averages presented in this study will be of more value for extension than the shorter period results such as the daily values. This is implied by the errors that have been estimated (section 1.5).

3. Values of SW↓ are higher for the study year (293 ly day^{-1}) than for the 1965-67 average (276 ly day^{-1}), and there were atypical values within the study year, notably in November and December. However, data from the Chilton Valley and the Craigieburn station suggest that de Lisle's (1966) maps of SW↓, provide a good approximation of the monthly values of this important parameter. They could be used to extend heat balance estimates to other parts of the High Country with the use of appropriate SW↓/Rn relationships and other information, and/or the use of climatonomy.

4. Precipitation for the study year (91 cm) was below the 47 year average for the Biological Station (131 cm).

5. Differences in aspect and slope lead to large variations of SW↓ (section 2.9) and consequently to large differences in energy exchange.

6. The discussion of the relationship between energy exchanges and synoptic scale weather situations (section 7.3) will be of value, as the latter, by definition, occur over a wide area. The application of such a study has been suggested by Holmes and Dingle (1965). This does not necessarily imply that the associations found at the

Chilton Valley apply to the remainder of the High Country.

7. Some of the factors discussed in this study have been shown to apply to other parts of the South Island. The overestimation of potential evapotranspiration, for example, by a Penman type formula has been found by Fitzgerald and Rickard (1960).

8. Finally, climatonomic theory appears to be a particularly potent tool for the extension of the present results and also in making new heat balance estimates, provided that, in both cases, care is taken in the selection of input data.

8.3 Suggestions for Further Research

At many points in this study, and especially in Chapters 2 - 5, the need for further research has been expressed. Some of the more important areas are summarised below.

1. There is a need for direct winter measurements of Rn. Methods of obtaining efficient unattended operation of radiation sensors in this season should be developed.

2. The sampling problem in the study of soil heat flow should be investigated in order to determine how representative are a small number of measurement positions.

3. The sampling problem in the measurement of soil moisture should also be studied.

4. The modified Penman model for obtaining daily values of actual evapotranspiration should be improved. Suggestions on how this can be done were given in section 4.4.h.

5. Since sensible and latent heat, and momentum fluxes are often treated by the same body of aerodynamic theory (Priestley, 1959), it is necessary to obtain more data on temperature, humidity, and wind profiles to investigate thoroughly whether the theory can be of value in rugged inhomogeneous terrain.

6. There is a need for the direct measurement of P, since even if the errors in daily values of Rn, LE and A (section 1.5) were halved, the resulting error in P, when taken as a residual, would still be about 13%.

7. The importance and influence of advection should be examined within the High Country environment.

8. The relationships between climatic and ecological energetics should be established.

9. A specific study on the spatial extension of results found in the Chilton Valley should be undertaken.

The research reported in this study is part of a detailed investigation of the physical geography of the Cass Basin (Soons and Rayner, 1968). The present study represents an initial examination of the heat balance, upon which, more accurate and sophisticated work can be based. The usefulness of the heat balance approach has been demonstrated. Subsequent research will further verify Geiger's (1965) statement that 'real understanding of microclimatological phenomena can be achieved only when all factors involved in heat balance are followed quantitatively throughout their sphere of influence'.

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APPENDIX A

MISSING DATA

Partial or complete loss of temperature, radiation, and soil heat flux data occurred on the following days:- 21 September, 19 October 1969; 15 February, 8, 22, 31 March, 5 April, 3 May, 7 June, 1970.

Radiation data were obtained for the missing days using the regression equation:

$$SW\downarrow \text{ (Chilton)} = 0.859\ SW\downarrow \text{ (Craig)} - 4.454 \text{ ly day}^{-1} \qquad \text{A.1.}$$

$$\text{C.C.} = 0.883 \qquad \text{S.E.E.} = \pm 94.9 \text{ ly day}^{-1}$$

Mean air temperatures were obtained from the regression equation:

$$T \text{ (Chilton)} = 0.909\ T \text{ (Craig)} + 1.077^{\circ}\text{C} \qquad \text{A.2.}$$

$$\text{C.C.} = 0.901 \qquad \text{S.E.E.} = \pm 2.2^{\circ}\text{C}$$

Soil heat flow data were obtained from 3.3.2.

Partial or complete loss of wind data occurred on the following dates:- 24 November - 2 December, 11-15 December, 22-28 December, 1969; 9-11 January, 21-25 February, 5-7, 10, 13-25 March, 29 March - 9 April, 15 May - 4 June, 1970. Data were obtained from the following equation:

$$u \text{ (Chilton)} = 0.761\ u \text{ (Craig)} + 1.785 \text{ m sec}^{-1} \qquad \text{A.3.}$$

$$\text{C.C.} = 0.586 \qquad \text{S.E.E.} = \pm 1.17 \text{ m sec}^{-1}$$

where u (Chilton) is the average wind velocity for the day in m sec^{-1}.

APPENDIX B

ESTIMATION OF ERRORS IN THE MAIN HEAT BALANCE COMPONENTS

1. Principles Used in Estimation

The main assumptions used in the estimation of the likely errors in the values of the heat balance components were given in section 1.5. In the following discussion two further rules are used. These are:-

(1) Combination of two or more errors is made by taking the square root of the sum of the squares of the errors. This rule can be applied to random errors where no products, quotients or powers, other than unity, are involved (Topping, 1955).

(2) Owing to the absence of other data, in the cases of SW↓ and Rn, a principle stated by Brooks and Carruthers (quoted by Stanhill, 1965) is employed. They suggest that the standard deviation of climatological measurements is inversely proportional to the square root of the length of the period considered.

2. Shortwave Radiation

An estimation of error in SW↓ is required for the subsequent calculation of error in Rn. SW↓ suffered from instrumental and sampling errors. Instrumental error for the pyranometer used, is usually quoted as 5% (Stanhill, 1965). de Vries (1955) also quotes the same error but suggests that part of this value is due to difficulties in obtaining integrated totals from the recorder chart. It will be assumed here, that errors attributed to such difficulties are included in the 5% instrumental error. A sampling error is introduced by taking readings at 7½ minute intervals instead of using continuous, or shorter time interval, records. In order to assess the possible magnitude of the sampling error, a continuous record from an Epply pyranometer at Christchurch Airport was examined. The record of a day with extremely variable cloud cover, and consequent high fluctuations of SW↓, was used to obtain daily totals by integrations of 7½ minute and 1 minute interval data. The latter interval was the shortest that could be obtained from the chart by eye. In this case, the error in using 7½ minute sampling, instead of 1 minute sampling, was approximately 8%. A further simplifying assumption is made; that the occurrence of days with different degrees of cloud cover variation, leading to variation in values of SW↓, is normally distributed between days with continuous cloud cover, or clear sky, at one extreme, to days with highly fluctuating cloud cover at the other. Assuming the error caused by sampling at 7½ minutes in fluctuating cloud cover conditions also to be normally distributed, the associated sampling error would then vary from 8% to 0% and be centred on 4%.

Combination of the 5% instrumental error, and the likely 4% sampling error, gives rise to a total probable error of 6.4% for a daily value. This represents 15.6 ly day^{-1} of the mean daily value for the study period. If this is assumed to be the standard deviation for the study year measurements, then the application of the principle of Brooks and Carruthers can be made (see Stanhill,1965). In this case, if the error for a daily value is 6.4%, then the error for a 30 day month is 1.2%, and that for the year is 0.3%.

3. Net Radiation

An average daily value of Rn is subject to the error in SW↓ (6.4%) and the error indicated by the value of the S.E.E. in equation 2.5.13 (31.9 ly day^{-1} or 23% for the mean daily value of Rn for the study year). Part of the latter error may be due to previously described errors in SW↓. If this is true, then to use the full 23% value in the following calculation may lead to an overestimate of error. However, since there is no method of determining how much error, in the S.E.E., is attributed to error in SW↓, the possibility of an overestimate must be accepted. Combination of the 6.4% and 23% errors gives an error of 23.8% for a daily value of Rn. Application of the Brooks and Carruthers principle results in an error of 4.4% for a monthly value and 1.3% for the annual value of Rn.

4. Soil Heat Flow

The absence of information on errors in soil heat flow measurement is conspicuous in many published reports of surface energy measurements. This is probably due to the relatively small size of the soil heat flow term, and the practical difficulty in estimating the error of this measurement. Uncertainties in the present case are discussed in sections 3.3 and 3.4. Difficulties in such estimation are reported by Fuchs and Tanner (1967), and possible reasons for inaccuracy are given by Phillip (1961). Frankenberger (1962), who used soil temperature records, lists possible causes of non-zero heat flow over an annual period. As mentioned in section 3.3, the manufacturer's calibration of the flux plates indicated an accuracy of 5%, which is similar to that found by Deacon (1950). However, owing to the general uncertainties, which include sampling errors, discussed here, and in other sections, a value of 5-10% for daily values is adopted for the present estimates. Again, due to uncertainty, no value is assumed for errors of monthly and annual values of soil heat flow.

5. Evaporative Heat Loss

In contrast with the literature on soil heat flow errors, that on the accuracy of evapotranspiration estimation, and thus evaporative heat loss, is vast. Works by Tanner and Pelton (1960) and McCaughey (1968) are important contributions with regard to the Penman Method. The latter found the Penman method to be accurate within 5% for daily values, when PET conditions were fulfilled. However, since the method used in the present study differs from the original Penman methods or method, it is thought more reasonable to follow the Budyko (1958) approach, of using the water balance to estimate the accuracy of the annual heat loss by evapotranspiration It is assumed that the water balance values are approximated by the lysimeter results. The latter are assumed to be correct, although possible errors are fully discussed in Appendix G. Comparison of annual totals of AET given by the lysimeters, and the modified Penman model, indicates a possible error of 10% for the annual value. The order of error for shorter periods may be obtained by a comparison of lysimeter and modified Penman model values, for the period January 24 - February 8 (see section 4.3). Such a comparison indicates a possible error of approximately 5%. In the absence of other data, this will be assumed to be the error for a monthly period. An estimate of error for daily values is given by assuming PET to have occurred, and combining the likely error in using a Penman formula (5% according to McCaughey), with the error due to using the modified model, instead of the actual Penman calculation. The latter error is estimated by applying the S.E.E. in equation 4.4.4., to the average daily evapotranspiration during the period for which the equation is applicable. This results in an error of 15%. Combination of the two types of error (5% and 15%) renders a value of 15.8% for the likely error in the daily values of E, and hence LE. The changing value of L with respect to temperature is neglected.

The estimated errors for daily and monthly values of LE should, strictly speaking, be applied only to the January and February data used in their estimation. It is assumed however that they apply to the whole year. This assumption may not be valid for the November period, when AET was less than PET.

6. Sensible Heat Flow

Since values of P are derived from closing the heat balance equation, estimates of error in the value of this term may be given by combining the errors in Rn, A and LE. For the purposes of this calculation only, errors in daily, monthly, and annual values of A are assumed to be 10, 1.8 and 0.5% respectively (Brooks and Carruther's principle having been applied to a daily value of 10%). Combination of the errors in Rn, A and LE results in an error of P of 29.7, 6.9 and 10.1% respectively for daily, monthly and annual values.

The values of error in LE and P are not in accord with the principle of Brooks and Carruthers. This is due to the employment of the water balance approach to obtain errors in E, and hence LE and P. Despite the discrepancy it was thought more appropriate to use, where possible, the more direct water balance approach, rather than the Brooks and Carruther's principle.

APPENDIX C

AN INTRODUCTION TO THE THEORY OF CLIMATONOMY

The study of the surface energy exchange in terms of forcing and response functions forms part of a large body of theory called climatonomy by its principal exponent, H. Lettau. Since only a small part of the theory of climatonomy has been published widely to date (Lettau, H. and Lettau, K., 1969; Lettau, 1969) an outline of the details relevant to the present study is given below. This is based on documents of restricted circulation (Lettau, 1968).

Net shortwave radiation is regarded as a forcing function, F, which is given by

$$F = (1 - \alpha)\, SW\downarrow \qquad \text{—— C.1.}$$

and, for the standard terrestial case, gives rise to a response function, F', given by the equation

$$F' = \varepsilon \sigma T_o^4 + LW\downarrow + A + P + LE \qquad \text{—— C.2.}$$

Since all the terms on the right hand side of equation C.2. can be related directly, or indirectly, to T_o, the latter itself may be regarded as the response function. Further, involving the principle of conservation of energy, the two right hand sides of equations C.1. and C.2. may be equated. Both those, and the resulting equation are instantaneously valid, and can be applied to a time series of instantaneous values formed by both the forcing and response functions. When viewed in terms of time series, consideration of the basic frequencies, and their higher harmonics, produces analogies to such fields as 'forced vibrations' in (1) consideration of amplitude ratios (impedances) and phase differences (laggings) between the forcing and response functions, and (2) dependency on frequencies.

In terms of cycles, the basic relationships of the theory are developed as follows. Let n_i, with $i = 1, 2, 3, \ldots$ denote the first and higher harmonics of the basic cycle of frequency n or period τ, where τ equals $2\pi/n$. Let an overbar indicate a time average between times t and $t + \tau$. With Fourier expansion, the forcing function becomes:

$$F(t) = \bar{F} + \sum_i \Delta_i F \cos (n_i t - \delta_i) \qquad \text{C.3.}$$

where Δ_i is the amplitude, $n_i t$ is the wave or harmonic number, δ_i is the phase angle, and the last term is summed over all the harmonics in the analysis. The response function becomes:

$$T_o(t) = \bar{T}_o + \sum_i \Delta_i T_o \cos (n_i t - \delta_i^*) \qquad \text{C.4.}$$

where δ_i^* is the phase angle of the response function. The total climatic impedance, Z_i, is defined as $\Delta_i F/\Delta_i T_o$. The total phase lag, ζ_i, is defined as $\delta_i^* - \delta_i$. The response function can be reformulated as a predictor equation as given by

$$T_o(t) = \bar{T}_o + \sum_i (\Delta_i F/Z_i) \cos (n_i t - \delta_i - \xi_i) \qquad \text{C.5.}$$

By considering the departure form of the forcing function equation,

$$F - \bar{F} = LW\uparrow - \overline{LW}\uparrow + LW\downarrow - \overline{LW}\downarrow + A - \bar{A} + P - \bar{P} + LE - \overline{LE} \quad \text{C.6.}$$

and i sets of partial impedances (Γ_i, β_i, Ψ_i, Φ_i, χ_i), and i corresponding sets of phase constants (γ_i, b_i, ψ_i, φ_i, x_i) for all n_i, it is possible to derive (see Appendix E) the following equations for Z_i and ζ_i.

$$Z_i = \Gamma_i \cos(\zeta_i - \gamma_i) + \beta_i \cos(\zeta_i - b_i) + \Psi \cos(\xi_i - \psi_i) + \Phi \cos(\xi_i - \varphi_i) + \chi_i \cos(\xi_i - x_i) \quad \text{C.7.}$$

$$\tan \zeta_i = (\Gamma_i \sin \gamma_i + \beta_i \sin b_i + \Psi_i \sin \psi_i + \Phi_i \sin \varphi_i + \chi_i \sin x_i)/(\Gamma_i \cos \gamma_i + \beta_i \cos b_i + \Psi_i \cos \psi_i + \Phi_i \cos \varphi_i + \chi_i \cos x_i) \quad \text{C.8.}$$

The Fourier analysis of F yields $\Delta_i F$ and δ_i for a variety of basic frequencies. The immediate problem is to obtain the sets of partial impedances and phase constants, by means of model assumptions and parameterisations (see Appendix E). In practice ζ_i of equation C.8. and then Z_i of equation C.7. are found, whereupon equation C.5. is used to compute surface temperature variations. The period mean surface temperature can be found from the period mean part of C.6. The time series of the energy budget components can be obtained when parameterisation (Appendix E) is complete.

APPENDIX D

THE COMPUTATION OF MEAN ANNUAL AIR TEMPERATURE FROM ESTIMATES OF NET LONGWAVE RADIATION

Data for the computation of the mean annual air temperature are shown in Table D.1. Values of relative humidity and mean vapour pressure are average values of the years 1964-67, at the Craigieburn Forest station (N.Z.M.S., 1966-70). The friction velocity at 2 m is computed from the equation:

$$u_* = (u_{200}\ k)/(\ln(200/z_o)) \qquad \text{D.1.}$$

where u_{200} is the windspeed at 2 m. (Haltiner and Martin, 1957, p.228). The Angstrom ratio, A^o, equals $(LW\uparrow + LW\downarrow)/LW\uparrow$. It is computed using

$$(1)\quad LW\downarrow = \varepsilon\sigma T^4\ (a_o - b_o\ 10^{-c_o e}) \qquad \text{D.2.}$$

(Sellers, 1965, p.53), where T is the air temperature near the surface, e is the vapour pressure in mm Hg, and the constants a_o, b_o, c_o are respectively 0.820, 0.250 and 0.094 (Geiger, 1965, p.20).

$$\text{and}\quad (2)\quad LW\uparrow = \varepsilon\sigma T^4 \qquad \text{D.3.}$$

Following Lettau (1968), to allow for cloudiness, $0.66A^o$ is used in the annual computations. The mean surface temperature $\overline{T}_o$ is obtained from

$$\overline{P} = A^o\ \varepsilon\sigma\overline{T}_o^{\,4} \qquad \text{D.4.}$$

where $\overline{P}$ is the mean annual upward sensible heat flow in m ly min^{-1}. The difference between the mean air temperature at the surface, $\overline{T}_o$, and that at 2 m, T_{200}, is obtained from

TABLE D.1.

DATA AND ASSUMPTIONS NEEDED FOR CHECKING THE MEAN ANNUAL HEAT BUDGET ESTIMATES

Chilton Valley mean annual air temperature during the study period	8.8°C
Relative Humidity (N.Z.M.S., 1966-70)	74%
Windspeed. Chilton Valley during the study year. (u_{300} is assumed to equal u_{200})	2.4 m sec^{-1}
Mean air density $\bar{\rho}$ at 800 m (C.O.E.S.A., 1962)	1.13 mg cm^{-3}
Specific heat of air at constant pressure (Hess, 1959)	0.24 cal.g^{-1} $°C^{-1}$
Mean vapour pressure (from N.Z.M.S. 1966-70, for Craigieburn)	6 mm Hg
Roughness length, z_o (Assumed as in section 4.4.f. see also section 5.4)	9.0 cm
Friction velocity, u_* (at 240 cm sec^{-1})	27.2 cm sec^{-1}
Surface emissivity, ε (Brooks, 1959)	0.96
Angstrom Ratio, A^o	0.25

the equation

$$\overline{T}_{200} - \overline{T}_o = -\overline{P} (\log_e 200 - \log_e z_o)/(0.428 \, c_p \overline{\rho} \, u_*) \qquad \text{D.5.}$$

which is given by Lettau (1968). This difference, which is usually negative, is then added to the derived value of $\overline{T}_o$, and results in the value of $\overline{T}_{200}$.

APPENDIX E

PARAMETERISATIONS AND THE BASIC RELATIONSHIPS OF CLIMATONOMY

The parameterisation formulae, quantitative assumptions, and values used, are listed below.

1. The caloric admittance (Lettau's term) of the submedium, μ , is given by

$$\mu = \sqrt{kC} \quad \text{m ly } {}^{o}C^{-1} \sec^{-\frac{1}{2}} \qquad \text{—— E.1.}$$

(Lettau, 1968), where k is taken to be 3.33 m ly sec^{-1} $^{o}C^{-1}$ cm^{-1}, which is the value for 4 October 1966 in Table 3.1, and C is taken as 0.55 cal $^{o}C^{-1}$ cm^{-3}, which is selected as being representative following the discussion in section 3.2.

2. The values of $c_p \bar{\rho}$, z_o, and u_* are taken from the data in Table D.1.

3. The phase angle of soil heat flow, ψ , is assumed to be 45^o following Lettau (1968).

4. In order to obtain the impedances Γ_i, β_i for the longwave radiatin the following procedure is adopted (Lettau,1968)

$$\Gamma = 4 \, \overline{LW}\uparrow / \overline{T}_o \qquad \text{—— E.2.}$$

and

$$\beta = \Gamma r^* \quad (A^o - 1) \qquad \text{—— E.3.}$$

where

$$r^* = e^{(-n/n^*)} \qquad \text{—— E.4.}$$

In the annual calculations $n^* = 18 \times 10^{-6}$, and the values of n from the first to the fourth harmonics respectively are 20, 40, 60 and 80, all $\times 10^{-8}$, giving values of r* of respectively 0.99, 0.98, 0.97 and 0.96. Adding Γ_i to both sides of equation E.3. gives

$$\Gamma_i + \beta_i = \Gamma_i (1 + r^* (A^o - 1)) \qquad \text{E.5.}$$

where subscript i refers to the harmonic number. Values of $\Gamma_i + \beta_i$ are computed from equation E.5. using equation E.2.

5. Values of Ψ_i are computed from

$$\Psi_i = \mu \sqrt{n_i} \qquad \text{E.6.}$$

6. Values of N_i are given by

$$N_i = \log_{10} (u_* / (n_i z_o)) \qquad \text{E.7.}$$

7. Values of Φ_i are calculated from

$$\Phi_i = 1.8 c_p \bar{\rho} u_*/(1 + N_i^{2.5}) \qquad \text{E.8.}$$

8. Values of φ_i are calculated from

$$\varphi_i = \cot^{-1} (1 + 0.020 N_i^3) \qquad \text{E.9.}$$

9. Parameterisations 5 to 8 all follow Lettau (1968). The same author also assumes that the phase angles of P, φ , and LE, x_i, are equal, and that the impedance of LE, χ_i is given by

$$\chi_i = B' \Phi \qquad \text{E.10.}$$

where B' is an adjusted value of the inverse Bowen ratio. The adjustment is necessary since LE cannot show large flows to the surface, as is possible with P. In the present calculation the inverse Bowen ratio, B, is not adjusted, and the effect on the results is discussed in section 6.3.c.

10. The equations for obtaining Z_i and ζ_i were given in Appendix C, as equations C.7. and C.8. Lettau (1968) derives these equations using the following explanation. It is noted that a function of the form $f(\theta) = \cos(nt - \delta - \xi + X)$ may be expanded as

$$\cos(nt - \delta - \xi + X) = \cos(nt - \delta) . \cos(\xi - X) + \sin(nt - \delta) \sin(\xi - X) \qquad \text{E.11.}$$

It is also noted that the departure form of the energy budget equation (equation C.6.), must be satisfied simultaneously for the sum of the factors of $\cos(nt - \delta)$, as well as $\sin(nt - \delta)$. Since the partial impedances, Γ_i, β_i, Ψ_i, Φ_i and χ_i, and the phase constants, γ_i, b_i, ψ_i, φ_i and x_i, are defined by the equations

$$(LW\uparrow - \overline{LW}\uparrow)/\Delta_i T_o = \Gamma_i \cos(n_i t - \delta_i - \xi_i + \gamma_i) \qquad \text{E.12.}$$

$$(LW\downarrow - \overline{LW}\downarrow)/\Delta_i T_o = \beta_i \cos(n_i t - \delta_i - \xi_i + b_i) \qquad \text{E.13.}$$

$$(A - \overline{A})/\Delta_i T_o = \Psi_i \cos(n_i t - \delta_i - \xi_i + \psi_i) \qquad \text{E.14.}$$

$$(P - \overline{P})/\Delta_i T_o = \Phi_i \cos(n_i t - \delta_i - \xi_i + \varphi_i) \qquad \text{E.15.}$$

$$(LE - \overline{LE})/\Delta_i T_o = \chi_i \cos(n_i t - \delta_i - \xi_i + x_i) \qquad \text{E.16.}$$

for all n_i equation C.6. yields the following two equations, which are time independent and exactly valid for any n_i, and where $Z_i = \Delta F/\Delta_i T_o$,

$$Z_i = \Gamma_i \cos(\xi_i - \gamma_i) + \beta_i \cos(\xi_i - b_i) + \Psi_i \cos(\xi_i - \psi_i) + \Phi_i \cos(\xi_i - \varphi_i) + \chi_i \cos(\xi_i - x_i) \qquad \text{E.17.}$$

which is also equation C.7., and

$$0 = \Gamma_i \sin(\xi - \gamma_i) + \beta_i \sin(\xi_i - b_i) + \Psi_i$$
$$\sin(\xi_i - \psi_i) + \Phi_i \sin(\xi_i - \varphi_i) +$$
$$\chi_i \sin(\xi_i - x_i) \qquad \text{——— E.18.}$$

Expansion of $\sin(\xi_i - X_i)$ is given by

$$\sin(\xi_i - X_i) = \sin \xi_i \cos X_i - \cos \xi_i \sin X_i \qquad \text{——— E.19.}$$

and thus equation E.18. yields

$$\tan \xi_i = (\Gamma_i \sin \gamma_i + \beta_i \sin b_i + \Psi \sin$$
$$\psi_i + \Phi_i \sin \varphi_i + \chi_i \sin x_i)/$$
$$(\Gamma_i \cos \gamma_i + \beta_i \cos b_i + \Psi \cos$$
$$\psi_i + \Phi_i \cos \varphi_i + \chi_i \cos x_i) \qquad \text{——— E.20.}$$

which is equation C.8.

The values of Z_i and ξ_i are thus obtained from equations E.17 and E.20 respectively, with the aid of the additional information, given by Lettau (1968), that

$\gamma_i = b_i = 0.$

11. The values of the amplitudes ΔT_o, $\Delta(LW\uparrow + LW\downarrow)$, ΔA, ΔP and ΔE are given by the following

$$\Delta T_{oi} = \Delta F_i / Z_i \qquad \text{——— E.21.}$$

$$\Delta(LW\uparrow + LW\downarrow)_i = \frac{\Delta F_i (\Gamma_i + \beta_i)}{Z_i} \qquad \text{——— E.22.}$$

$$\Delta A_i = \Psi_i \frac{\Delta Fi}{Z_i} \qquad \text{——— E.23.}$$

$$\Delta P_i = \Phi_i \frac{\Delta Fi}{Z_i} \qquad \text{——— E.24.}$$

$$\Delta LE_i = B \, \Delta P_i \qquad \text{——— E.25.}$$

12. The values of the response parameters that vary with each harmonic for the analysis in which the Bowen ratio is taken as 0.2 are shown in Table E.1.

13. An equation for the Fourier synthesis of ($LW\uparrow + LW\downarrow$) can be obtained by adding equations E.12 and E.13, and rearranging to give

$$(LW\uparrow + LW\downarrow) = (\overline{LW}\uparrow + \overline{LW}\downarrow) + \Delta_i T_{oi} (\Gamma_i + \beta_i) \cos (n_i t - \delta_i - \xi_i + \gamma_i) \quad \text{E.26.}$$

since $\alpha = \beta = 0$.

14. The mean terms in the Fourier synthesis were taken as the mean terms in the original data (Tables 6.1 and 6.11).

TABLE E.1.

VALUES OF RESPONSE PARAMETERS THAT VARY WITH THE FIRST TO THE FOURTH HARMONICS. IMPEDANCES, AND AMPLITUDES OF THE HEAT BALANCE COMPONENTS ARE IN M LY SEC^{-1}. PHASE ANGLES ARE IN DEGREES, AND TEMPERATURE AMPLITUDES ARE IN ^{o}C. CALCULATION AS FOR TABLE 6.12 BUT WITH BOWEN RATIO = 0.20

Harmonic	$\Gamma + \beta$	Ψ	N	Φ
1	0.02088	0.00630	7.1793	0.09511
2	0.02184	0.00896	6.8781	0.10560
3	0.02279	0.01092	6.7022	0.11250
4	0.02375	0.01260	6.5771	0.11820

Harmonic	φ	ξ	Z	ΔT_o
1	6.8	6.9	0.5985	3.943
2	7.6	8.0	0.6623	0.148
3	8.1	8.4	0.7063	0.167
4	8.5	8.8	0.7428	0.384

Harmonic	Δ(LW↑ + LW↓)	ΔA	Δ P	ΔLE
1	0.0823	0.025	0.375	1.875
2	0.0032	0.001	0.016	0.080
3	0.0038	0.002	0.019	0.095
4	0.0091	0.005	0.045	0.225

APPENDIX F

DAILY HEAT BALANCE DATA AND MEAN AIR TEMPERATURE

Heat balance data in ly day^{-1}, positive values indicating flow towards the surface. Air temperature in ^{o}C.

DATE	NET RADIATION	SOIL HEAT FLOW	EVAPORATIVE HEAT FLOW	SENSIBLE HEAT FLOW	AIR TEMP
15/ 8/69	35	6	-19	-22	4.0
16/ 8/69	67	16	-39	-45	2.7
17/ 8/69	46	15	-29	-33	2.5
18/ 8/69	99	20	-55	-65	2.7
19/ 8/69	92	21	-52	-61	2.7
20/ 8/69	58	14	-38	-34	0.7
21/ 8/69	1	22	-11	-13	-1.3
22/ 8/69	13	20	-22	-11	1.3
23/ 8/69	16	23	-19	-21	1.7
24/ 8/69	70	18	-41	-47	0.5
25/ 8/69	119	20	-64	-75	2.5
26/ 8/69	127	22	-76	-73	4.0
27/ 8/69	61	13	-47	-27	3.0
28/ 8/69	112	22	-65	-68	3.6
29/ 8/69	132	20	-73	-79	3.3
30/ 8/69	121	19	-70	-70	4.2
31/ 8/69	124	13	-104	-33	7.9

DATE	NET RADIATION	SOIL HEAT FLOW	EVAPORATIVE HEAT FLOW	SENSIBLE HEAT FLOW	AIR TEMP
1/ 9/69	144	17	-113	-48	6.3
2/ 9/69	145	20	-89	-76	5.0
3/ 9/69	25	18	-52	9	7.7
4/ 9/69	34	22	-45	-11	7.0
5/ 9/69	69	23	-53	-39	8.0
6/ 9/69	47	28	-51	-25	8.5
7/ 9/69	25	28	-45	-7	11.1
8/ 9/69	32	40	-53	-19	10.0
9/ 9/69	94	35	-75	-54	10.3
10/ 9/69	56	35	-57	-35	10.9
11/ 9/69	79	20	-62	-37	7.9
12/ 9/69	43	20	-44	-19	8.7
13/ 9/69	27	21	-31	-18	4.0
14/ 9/69	180	24	-100	-104	3.0
15/ 9/69	22	21	-25	-18	4.8
16/ 9/69	102	18	-71	-49	5.5
17/ 9/69	128	17	-75	-70	4.9
18/ 9/69	217	21	-135	-103	4.9
19/ 9/69	189	29	-104	-114	2.3
20/ 9/69	80	19	-56	-43	4.2
21/ 9/69	193	20	-121	-92	5.3
22/ 9/69	212	25	-137	-101	7.3
23/ 9/69	217	20	-134	-102	6.9
24/ 9/69	113	15	-71	-57	6.8
25/ 9/69	61	14	-40	-35	6.3
26/ 9/69	59	14	-46	-28	7.8
27/ 9/69	238	19	-144	-112	8.5
28/ 9/69	160	17	-135	-42	9.0
29/ 9/69	163	15	-106	-71	6.8
30/ 9/69	221	20	-130	-111	7.4

DATE	NET RADIATION	SOIL HEAT FLOW	EVAPORATIVE HEAT FLOW	SENSIBLE HEAT FLOW	AIR TEMP
1/10/69	242	14	-142	-114	8.7
2/10/69	101	20	-52	-69	3.5
3/10/69	95	22	-50	-67	0.7
4/10/69	258	10	-167	-102	5.7
5/10/69	269	3	-164	-108	6.5
6/10/69	63	-1	-28	-34	3.8
7/10/69	81	-2	-43	-36	6.1
8/10/69	98	-8	-47	-42	7.0
9/10/69	248	-1	-129	-118	5.2
10/10/69	51	11	-37	-26	-2.0
11/10/69	17	0	-9	-8	1.8
12/10/69	70	-3	-60	-6	6.0
13/10/69	281	-7	-160	-115	8.7
14/10/69	255	-10	-140	-105	10.6
15/10/69	284	-10	-172	-103	10.2
16/10/69	289	-3	-150	-136	8.8
17/10/69	114	-1	-104	-9	6.5
18/10/69	220	-8	-121	-91	8.4
19/10/69	277	-5	-171	-101	8.9
20/10/69	300	-5	-173	-121	7.2
21/10/69	307	5	-159	-153	6.3
22/10/69	83	1	-101	17	6.7
23/10/69	244	-5	-170	-70	7.2
24/10/69	183	-6	-98	-79	6.8
25/10/69	295	-12	-191	-92	9.4
26/10/69	314	-4	-162	-149	8.8
27/10/69	132	2	-58	-75	1.1
28/10/69	100	-9	-46	-45	4.9
29/10/69	321	-8	-165	-149	7.4
30/10/69	257	-16	-132	-110	7.8
31/10/69	281	-8	-142	-131	5.2

DATE	NET RADIATION	SOIL HEAT FLOW	EVAPORATIVE HEAT FLOW	SENSIBLE HEAT FLOW	AIR TEMP
1/11/69	142	-14	-70	-58	6.1
2/11/69	316	-19	-142	-155	10.1
3/11/69	327	-28	-175	-124	11.6
4/11/69	324	-18	-187	-119	12.1
5/11/69	329	-26	-126	-177	13.8
6/11/69	334	-21	-185	-128	14.4
7/11/69	343	-21	-198	-124	15.2
8/11/69	284	-19	-129	-137	12.0
9/11/69	252	-16	-113	-123	15.0
10/11/69	110	-8	-80	-22	12.2
11/11/69	278	-11	-86	-181	13.5
12/11/69	141	-5	-41	-95	13.0
13/11/69	356	-23	-76	-258	14.9
14/11/69	266	-13	-47	-206	14.9
15/11/69	296	-10	-41	-245	12.8
16/11/69	349	-9	-52	-288	17.2
17/11/69	357	-19	-55	-283	18.0
18/11/69	335	-12	-49	-273	16.1
19/11/69	310	-13	-45	-252	16.2
20/11/69	380	-8	-52	-320	14.0
21/11/69	72	1	-47	-26	12.5
22/11/69	118	2	-57	-62	6.7
23/11/69	159	10	-70	-99	4.1
24/11/69	380	-16	-173	-191	7.2
25/11/69	395	-22	-179	-194	10.4
26/11/69	201	-5	-108	-89	11.7
27/11/69	187	-1	-105	-82	12.8
28/11/69	385	-24	-200	-162	13.5
29/11/69	289	-27	-155	-107	15.3
30/11/69	50	-8	-25	-17	15.4

DATE	NET RADIATION	SOIL HEAT FLOW	EVAPORATIVE HEAT FLOW	SENSIBLE HEAT FLOW	AIR TEMP
1/12/69	162	-3	-83	-76	13.3
2/12/69	315	-7	-159	-150	14.9
3/12/69	347	-13	-196	-137	16.6
4/12/69	171	-4	-93	-74	12.6
5/12/69	354	-10	-177	-167	10.7
6/12/69	323	-13	-183	-127	11.2
7/12/69	231	-5	-128	-99	12.0
8/12/69	101	2	-60	-43	12.3
9/12/69	339	-10	-180	-148	11.4
10/12/69	255	-12	-136	-107	12.8
11/12/69	193	-2	-107	-84	14.0
12/12/69	208	-2	-101	-105	8.6
13/12/69	190	-4	-91	-94	10.7
14/12/69	263	-12	-140	-111	13.8
15/12/69	310	-11	-177	-122	16.5
16/12/69	209	-11	-109	-89	14.9
17/12/69	38	4	-20	-21	9.3
18/12/69	156	-5	-83	-68	11.3
19/12/69	220	-12	-115	-94	13.7
20/12/69	316	-22	-173	-120	15.5
21/12/69	298	-6	-186	-107	17.6
22/12/69	393	-25	-217	-151	15.8
23/12/69	350	-18	-195	-138	15.7
24/12/69	147	-3	-84	-59	15.9
25/12/69	93	0	-45	-48	10.4
26/12/69	161	0	-77	-84	9.2
27/12/69	354	-12	-165	-177	10.8
28/12/69	306	-12	-163	-131	12.5
29/12/69	375	-26	-213	-136	16.6
30/12/69	307	-20	-176	-111	17.4
31/12/69	229	-5	-142	-83	18.0

DATE	NET RADIATION	SOIL HEAT FLOW	EVAPORATIVE HEAT FLOW	SENSIBLE HEAT FLOW	AIR TEMP
1/ 1/70	320	-22	-191	-107	17.9
2/ 1/70	335	-21	-185	-128	17.0
3/ 1/70	323	-24	-193	-105	20.8
4/ 1/70	275	-17	-152	-106	19.5
5/ 1/70	276	-16	-160	-100	18.5
6/ 1/70	162	1	-98	-65	16.4
7/ 1/70	115	6	-61	-60	6.5
8/ 1/70	337	-14	-156	-166	9.3
9/ 1/70	216	0	-119	-98	11.7
10/ 1/70	384	-22	-201	-161	14.6
11/ 1/70	342	-17	-180	-144	13.0
12/ 1/70	372	-17	-226	-129	15.8
13/ 1/70	163	-23	-99	-40	16.5
14/ 1/70	389	-25	-219	-145	16.6
15/ 1/70	363	-19	-191	-154	13.9
16/ 1/70	364	-22	-206	-136	15.8
17/ 1/70	338	-19	-200	-119	16.1
18/ 1/70	204	-9	-121	-74	13.9
19/ 1/70	274	-4	-143	-127	13.0
20/ 1/70	316	-19	-172	-126	14.2
21/ 1/70	176	-12	-75	-89	14.1
22/ 1/70	78	-5	-40	-32	14.2
23/ 1/70	71	-1	-39	-31	12.7
24/ 1/70	315	-2	-194	-119	17.8
25/ 1/70	206	-11	-134	-62	19.5
26/ 1/70	194	-9	-128	-57	18.7
27/ 1/70	303	-19	-181	-103	19.3
28/ 1/70	226	-4	-168	-54	19.0
29/ 1/70	121	-5	-64	-52	8.7
30/ 1/70	313	-6	-178	-130	10.1
31/ 1/70	328	-13	-201	-115	13.1

DATE	NET RADIATION	SOIL HEAT FLOW	EVAPORATIVE HEAT FLOW	SENSIBLE HEAT FLOW	AIR TEMP
1/ 2/70	340	-14	-210	-116	15.5
2/ 2/70	72	10	-65	-16	9.1
3/ 2/70	295	4	-168	-131	11.7
4/ 2/70	311	7	-179	-139	11.7
5/ 2/70	142	6	-126	-22	13.4
6/ 2/70	241	-5	-190	-46	19.1
7/ 2/70	177	0	-110	-67	16.2
8/ 2/70	112	16	-61	-68	4.3
9/ 2/70	269	7	-145	-130	8.0
10/ 2/70	316	-6	-188	-122	14.2
11/ 2/70	102	6	-79	-29	11.6
12/ 2/70	303	-8	-172	-124	13.1
13/ 2/70	234	6	-137	-103	11.9
14/ 2/70	83	19	-62	-40	7.6
15/ 2/70	251	13	-133	-131	5.7
16/ 2/70	300	-3	-169	-128	13.7
17/ 2/70	307	-2	-185	-120	15.7
18/ 2/70	297	5	-190	-111	16.4
19/ 2/70	270	-1	-194	-75	16.7
20/ 2/70	213	2	-182	-33	17.5
21/ 2/70	215	10	-132	-92	15.6
22/ 2/70	269	10	-156	-123	13.8
23/ 2/70	276	4	-157	-123	14.5
24/ 2/70	247	0	-147	-100	15.8
25/ 2/70	92	-1	-54	-37	12.9
26/ 2/70	267	-5	-165	-97	13.2
27/ 2/70	156	-2	-89	-65	13.3
28/ 2/70	120	-4	-76	-40	15.2

DATE	NET RADIATION	SOIL HEAT FLOW	EVAPORATIVE HEAT FLOW	SENSIBLE HEAT FLOW	AIR TEMP
1/ 3/70	152	2	-83	-71	9.4
2/ 3/70	270	-1	-164	-105	12.0
3/ 3/70	272	-2	-161	-109	14.3
4/ 3/70	270	-2	-155	-113	14.9
5/ 3/70	274	-2	-161	-111	15.6
6/ 3/70	66	-4	-35	-27	14.9
7/ 3/70	39	-2	-21	-16	11.7
8/ 3/70	140	-1	-79	-60	13.7
9/ 3/70	166	-3	-97	-67	17.4
10/ 3/70	147	-2	-85	-60	17.0
11/ 3/70	221	6	-147	-80	16.9
12/ 3/70	128	1	-107	-23	15.9
13/ 3/70	163	-3	-94	-65	16.7
14/ 3/70	29	12	-21	-21	8.8
15/ 3/70	96	15	-56	-54	5.3
16/ 3/70	31	5	-18	-18	6.9
17/ 3/70	163	3	-80	-87	8.6
18/ 3/70	147	6	-75	-77	8.3
19/ 3/70	145	2	-84	-63	11.3
20/ 3/70	109	1	-55	-55	10.3
21/ 3/70	64	6	-34	-36	7.1
22/ 3/70	80	9	-46	-43	7.9
23/ 3/70	158	5	-80	-83	10.5
24/ 3/70	92	1	-52	-42	11.8
25/ 3/70	94	0	-54	-40	12.5
26/ 3/70	20	6	-18	-7	11.3
27/ 3/70	164	5	-109	-61	12.9
28/ 3/70	142	8	-108	-42	13.0
29/ 3/70	85	6	-53	-38	12.8
30/ 3/70	31	6	-22	-15	14.1
31/ 3/70	57	13	-34	-35	10.5

DATE	NET RADIATION	SOIL HEAT FLOW	EVAPORATIVE HEAT FLOW	SENSIBLE HEAT FLOW	AIR TEMP
1/ 4/70	57	13	-38	-32	5.8
2/ 4/70	97	11	-58	-51	5.8
3/ 4/70	51	3	-34	-21	11.9
4/ 4/70	71	6	-48	-29	13.2
5/ 4/70	161	1	-102	-60	13.2
6/ 4/70	5	10	-14	-1	10.9
7/ 4/70	11	12	-17	-6	8.4
8/ 4/70	14	10	-14	-10	8.4
9/ 4/70	26	12	-27	-11	9.8
10/ 4/70	25	7	-39	6	10.7
11/ 4/70	143	6	-179	30	11.8
12/ 4/70	131	8	-113	-26	9.4
13/ 4/70	99	5	-76	-28	12.7
14/ 4/70	141	9	-130	-21	11.5
15/ 4/70	123	8	-157	26	11.7
16/ 4/70	82	8	-107	18	12.3
17/ 4/70	132	9	-129	-11	11.4
18/ 4/70	49	6	-66	11	14.0
19/ 4/70	132	8	-85	-55	8.6
20/ 4/70	110	12	-96	-26	5.1
21/ 4/70	117	11	-101	-27	7.1
22/ 4/70	116	8	-93	-31	10.1
23/ 4/70	97	6	-123	20	11.7
24/ 4/70	15	0	-17	3	11.8
25/ 4/70	34	-1	-40	7	12.0
26/ 4/70	103	2	-67	-37	10.8
27/ 4/70	50	5	-66	11	8.3
28/ 4/70	99	5	-95	-9	7.6
29/ 4/70	90	5	-93	-2	8.1
30/ 4/70	15	0	-12	-3	6.9

DATE	NET RADIATION	SOIL HEAT FLOW	EVAPORATIVE HEAT FLOW	SENSIBLE HEAT FLOW	AIR TEMP
1/ 5/70	28	3	-38	6	5.8
2/ 5/70	65	5	-47	-23	4.3
3/ 5/70	159	8	-80	-88	0.4
4/ 5/70	79	9	-45	-44	0.0
5/ 5/70	78	8	-44	-42	0.0
6/ 5/70	81	5	-45	-41	3.2
7/ 5/70	72	6	-38	-39	2.7
8/ 5/70	25	4	-20	-9	4.9
9/ 5/70	25	-2	-15	-8	6.7
10/ 5/70	43	-1	-35	-7	2.0
11/ 5/70	67	7	-45	-29	1.4
12/ 5/70	71	8	-44	-35	2.0
13/ 5/70	47	0	-57	9	7.0
14/ 5/70	56	5	-32	-29	3.5
15/ 5/70	55	4	-28	-32	3.5
16/ 5/70	54	2	-33	-23	7.5
17/ 5/70	53	2	-32	-22	10.2
18/ 5/70	17	-5	-12	0	9.4
19/ 5/70	48	0	-26	-22	7.9
20/ 5/70	25	-2	-13	-10	5.5
21/ 5/70	26	-4	-12	-10	7.8
22/ 5/70	24	-2	-11	-11	3.7
23/ 5/70	52	5	-29	-28	0.8
24/ 5/70	7	3	-6	-5	2.1
25/ 5/70	41	6	-22	-25	0.8
26/ 5/70	25	4	-14	-15	0.1
27/ 5/70	20	3	-11	-11	1.9
28/ 5/70	36	5	-19	-22	1.0
29/ 5/70	25	9	-16	-18	1.2
30/ 5/70	45	6	-24	-27	2.4
31/ 5/70	30	2	-18	-14	4.7

DATE	NET RADIATION	SOIL HEAT FLOW	EVAPORATIVE HEAT FLOW	SENSIBLE HEAT FLOW	AIR TEMP
1/ 6/70	32	-1	-20	-12	6.0
2/ 6/70	3	-4	0	1	6.0
3/ 6/70	9	-6	-4	1	8.5
4/ 6/70	6	6	-8	-5	9.9
5/ 6/70	-4	-1	0	5	7.8
6/ 6/70	19	3	-27	4	2.7
7/ 6/70	58	8	-34	-31	-1.2
8/ 6/70	11	7	-12	-6	-1.3
9/ 6/70	12	5	-11	-6	-2.0
10/ 6/70	3	10	-9	-4	-3.5
11/ 6/70	7	5	-10	-3	-0.8
12/ 6/70	-3	6	-4	1	-0.4
13/ 6/70	-5	5	0	0	1.0
14/ 6/70	20	-4	-19	3	6.0
15/ 6/70	8	-8	0	0	7.6
16/ 6/70	21	-3	-22	4	6.7
17/ 6/70	1	-1	0	0	3.6
18/ 6/70	0	1	-2	0	2.6
19/ 6/70	18	7	-18	-7	2.2
20/ 6/70	-12	6	0	6	-0.1
21/ 6/70	12	-2	-12	2	5.2
22/ 6/70	-8	5	0	3	1.5
23/ 6/70	4	0	-4	1	6.5
24/ 6/70	-1	2	-1	0	5.8
25/ 6/70	-9	4	0	5	4.4
26/ 6/70	-10	5	0	5	3.1
27/ 6/70	4	2	-7	1	9.1
28/ 6/70	26	-1	-30	5	8.4
29/ 6/70	14	-2	-15	2	6.6
30/ 6/70	19	1	-13	-7	3.6

DATE	NET RADIATION	SOIL HEAT FLOW	EVAPORATIVE HEAT FLOW	SENSIBLE HEAT FLOW	AIR TEMP
1/ 7/70	19	2	-25	4	5.6
2/ 7/70	18	0	-21	4	7.2
3/ 7/70	-10	1	0	9	3.4
4/ 7/70	0	2	-2	0	3.8
5/ 7/70	6	5	-13	2	5.4
6/ 7/70	20	3	-28	5	5.8
7/ 7/70	12	2	-8	-6	3.4
8/ 7/70	15	7	-18	-4	0.0
9/ 7/70	21	5	-31	5	4.8
10/ 7/70	15	-1	-17	3	5.8
11/ 7/70	4	0	-4	1	6.1
12/ 7/70	21	2	-21	-2	5.4
13/ 7/70	14	4	-12	-7	2.0
14/ 7/70	6	5	-10	-2	0.2
15/ 7/70	17	6	-16	-7	-1.5
16/ 7/70	38	7	-28	-18	-1.7
17/ 7/70	36	6	-32	-10	2.4
18/ 7/70	64	3	-80	13	5.0
19/ 7/70	50	0	-61	10	7.0
20/ 7/70	-7	1	0	6	5.7
21/ 7/70	12	7	-23	4	1.1
22/ 7/70	45	6	-42	-9	3.0
23/ 7/70	6	4	-12	2	3.4
24/ 7/70	26	5	-28	-3	3.0
25/ 7/70	14	-1	-16	3	8.1
26/ 7/70	38	2	-47	8	7.7
27/ 7/70	51	6	-36	-21	4.5
28/ 7/70	24	7	-20	-10	0.1
29/ 7/70	57	8	-41	-24	1.5
30/ 7/70	48	8	-31	-24	2.0
31/ 7/70	62	6	-42	-26	3.8

DATE	NET RADIATION	SOIL HEAT FLOW	EVAPORATIVE HEAT FLOW	SENSIBLE HEAT FLOW	AIR TEMP
1/ 8/70	23	6	-34	6	5.4
2/ 8/70	-11	1	0	10	6.8
3/ 8/70	28	4	-39	6	6.9
4/ 8/70	12	1	-15	2	6.5
5/ 8/70	31	6	-45	7	8.2
6/ 8/70	25	4	-35	6	10.7
7/ 8/70	67	10	-93	15	8.6
8/ 8/70	72	16	-106	18	8.5
9/ 8/70	73	10	-100	17	13.6
10/ 8/70	78	5	-99	17	13.2
11/ 8/70	84	4	-69	-20	13.8
12/ 8/70	82	7	-58	-31	14.2
13/ 8/70	25	1	-18	-7	16.0
14/ 8/70	101	9	-77	-33	12.5
15/ 8/70	92	12	-73	-31	11.9
16/ 8/70	97	13	-69	-40	9.4

APPENDIX G

POSSIBLE ERRORS IN THE LYSIMETER MEASUREMENTS

The purpose of this appendix is to examine some of the possible errors associated with the measurements made by the lysimeters, and the effect of these errors on the computation of values of AET. The errors are classified as (1) errors that are assumed to be random, and (2) errors that are partially or wholly systematic.

1. Random Errors

Random errors arise from the measurement of the surface catchment area of the field tanks and raingauges, and the volumes of water collected by these instruments. Ten separate measurements of the areas of the raingauges and field tanks showed standard deviations of 0.71 cm^2 and 25.9 cm^2 respectively. If these standard deviations are used to compute the fractional error in the areas, the fractional errors are 0.0037 for the raingauge and 0.0114 for the field tank. The average fortnightly totals of rain and percolated water are 1.06 cm and 0.52 cm respectively. On the basis of these totals, the fractional errors in the measured volumes of rain and percolated water are respectively 0.016 and 0.008, since the volumes could be measured to $\pm$ 3 and $\pm$ 10 cm^3. The two latter values are assumed; the assumptions being made on the basis of experience in the use of the measuring cylinder. No confidence limits are attached to these values.

The annual value of AET is calculated from the equation:

$$AET = (V_r/A_r) - (V_p/A_p) \qquad \text{G.1.}$$

where V_r and V_p are the volumes of rain and percolated water and A_r and A_p are the areas of the raingauge and field tank. Following rules given by Topping (1955), the fractional errors in the quotients, V_r/A_r and V_p/A_p, are 0.0123 and 0.0034 cm respectively. Combination of these fractional errors, by the rule given in Appendix B, results in a fractional error in the sum, $V_r/A_r \pm V_p/A_p$, of 0.01. On this basis, if the above assumptions are valid, errors in the annual value of AET, owing to instrumental measurement errors, amount to approximately one percent of the annual total.

2. Systematic Errors

Other errors to which the lysimeters may be subject are considered to be partly or wholly systematic. These are as follows:-

(1) The first is due to the difference in the water balance equations of the natural soil surface and the lysimeter monolith. The equation for the natural surface and soil column is

$$Ra - E - \Delta S - Pe - Ru = 0 \qquad \text{G.2.}$$

where Ra is precipitation, E is evaporation, ΔS is change in soil moisture storage, Pe is water which percolates out of the column. The equation for the column of soil in the field tanks is:

$$Ra - E - \Delta S - Pe = 0 \qquad \text{\_\_\_\_ G.3.}$$

since the sides of the field tank allow no runoff. Water that would otherwise have runoff is kept in the field tank, and affects the values of the E, ΔS, and Pe. The extent to which the values of these three terms individually are influenced cannot be determined with the present available data. However, Soons (1970) has shown that runoff in the Chilton Valley is usually less than 3% of incoming rainfall. Therefore the effect of the incorporated runoff will be small in this location.

(2) Another factor is the different sizes of the raingauge and lysimeter catchment areas. No independent tests were conducted on this, but Hamilton (1954) found that, for a slope with an aspect facing the prevailing storm direction, a catchment gauge of 28400 cm^2 at ground level but protected from rainsplash, collected 6.8% more rain than a gauge of 507 cm^2 area at 104 cm above the ground. Adjustment of the data for the different heights by means of a relationship of de Zeeuw (1963), indicates that the two gauges would have recorded the same amounts to within less than one percent. Therefore, there is some indication that differences in catch between the field tank and the raingauge, owing to differences in the areas of the two, will be less than one percent.

(3) If the vegetation in the lysimeters protrudes significantly

above the field tank, there is the possibility of an increased catch of precipitation owing to interception (Rowley, 1970). This is not believed to be a great problem in the present case since the plants in lysimeters 1-3 were relatively small. It is possible that interception gave rise to an increased catch in lysimeter 4 which contained a single large tussock. However, the effect was at least partly offset, in this case, by the relatively sheltered exposure of the lysimeter. If it is assumed that any increased catch goes into the percolation term, and allowing the maximum increase of catch found by Rowley (26%), then the annual value of AET measured by the lysimeters would be decreased by 3.4 cm. However, the above assumption is only correct for heavy rainfalls. In light rainfalls, when no percolation occurs, and when the soil is not at field capacity, the increased catch would go into the soil moisture and, at least in part, become available for ET. Therefore, it is possible that interception of rainfall by the vegetation in the field tanks, if occurring, will decrease the measured AET by a quantity less than 3.4 cm.

(4) In-splash, in all probability, did not occur in the raingauges since these, although at the same level as the lysimeter, were on the downhill side of the field tanks, and about 30 cm above the surface. The same is true of the downhill side of the field tanks, but some in-splash may have occurred on the uphill side,

since, at this point, the sides of the tank were between 10-15 cm above the surface. If in-splash occurred it would lead to a decrease in measured AET, if the same argument, as applied to point 3, is followed.

(5) Since the field tanks were not very deep, gypsum soil moisture blocks were inserted at 10 cm and 20 cm inside lysimeter 2, and just outside it at the same levels. This was in order to investigate whether the presence of the tanks disturbed the water retention properties of the soil. When the blocks were calibrated among themselves, the blocks which were later placed inside the field tanks, (C and D) showed values of percentage available soil moisture, approximately 3% higher than the blocks later placed outside the tank (Table 4.8). Even allowing for this, daily measurements of percentage available soil moisture in summer (Fig. 4.12) indicate that the available soil moisture inside the tank was about 6% greater than that outside the tank, during periods of no apparent soil moisture deficit. Further, when soil moisture at the 20 cm level begins to decrease, Fig. 4.12 shows that the soil moisture outside the tank is the first value to decrease. These observations are not conclusive because of the small number of calibration tests (Table 4.8), and the small number of blocks in the sample. However, it is possible that the field tanks stored moisture that would otherwise have percolated to greater depths. The observed slightly better state of health of the vegetation within the tanks tends to confirm this. If this is so, then the increased soil moisture would become

available for ET which would therefore be overestimated.

(6) Errors involved with the difficulties concerned with the measurement of snowfall are neglected, since there was very little snowfall during the study period. These difficulties include the catch of snowfall by both the raingauge and the field tank, and the blowing away of snow lying on the surface of the field tank.

(7) Errors associated with loss of water by evaporation from the raingauge are also neglected, since the water container of the raingauge was remote from the orifice of the gauge itself.

Overall, it is clear that a systematic quantitative study of the possible errors described above would be valuable. Relevant to the present study are the facts that some of the systematic errors work in opposite directions and/or are small. Errors (1) and (2) are small, and errors (3) and (4) will probably lead to underestimates in the value of AET whereas error (5) will lead to an overestimation of this term.